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A NORTHERN PERSPECTIVE

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A SOUTHERN PERSPECTIVE

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PREFACE[1]

UNESCO's International Hydrological Programme (IHP) launched the International Sediment Initiative (ISI) in 2002, taking into consideration that sediment production and transport processes are not sufficiently understood for practical uses in sediment management. Since information on ongoing research is an important support to sediment management, and bearing in mind the unequal level of scientific knowledge about various aspects of erosion and sediment phenomena at the global scale, a major mission of the ISI is to review erosion and sedimentation-related research. The two papers below were prepared in conformity with this important task of the ISI, following the decision of the ISI Steering Committee at its session in March 2004.

The subject of both papers is the modelling and prediction of sediment processes in watersheds and watercourses, which is essential for the development of sediment management policies and strategies in the respective regions. Since the tasks of sediment management differ from region to region, it is reasonable to discuss the problems and methods that prevail in one or another region in separate reviews. For this reason, the two papers published below are entitled: Erosion and sediment dynamics from catchment to coast, "A Northern Perspective" (by Professor Giampaolo DiSilvio from the University of Padua, Italy), and "A Southern Perspective" (by Professor Gerrit Basson, from the University of Stellenbosch, South Africa).

The differences between the northern and the southern perspectives indicate the different kinds of problems that sediment research has to deal with in different climatic and socio-economic conditions. Thus, the northern perspective essentially deals with problems encountered in Europe and Northern America, such as mobility of river sediments and their deposition in reservoirs and coastal zones, whereas the southern perspective puts the emphasis on the origin and mobility of the sediments in semi-arid zones and their impacts on land and water resources. However, it should be underlined that the concepts and methods described in the papers of either perspective can be applied also to other regions as appropriate. Generally, in any region, the most efficient methods of modelling would depend upon the nature of the problems that sediment management has to resolve in the concrete case. Accordingly, in subsequent papers on the same subject other aspects of erosion and sediment transport might also be examined which are relevant to typical management situations that may occur in different geographical and socio-economic settings.

It should be underlined that, in line with its general strategy, ISI is open to collaboration with international, regional, or national associations and institutions involved in the promotion of sustainable sediment management policies. ISI focuses on international information exchange on sediment-related matters, ensuring access to policy makers in UNESCO's Member States, and encouraging sediment research in interested regions and states. The publication of the two papers on erosion and sediment dynamics should, among others, serve this purpose. Readers are thus invited to react and contribute by way of comments and suggestions to the follow-up of these two reports.

[1] by Stevan Bruk - s.bruk@unesco.org

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A NORTHERN PERSPECTIVE: EROSION AND SEDIMENT DYNAMICS FROM CATCHMENT TO COAST

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1. BASIC FORMS OF SEDIMENT MOTION

Removal of sediment from the watershed slopes (erosion) and the subsequent discontinuous motion (dynamics) to the ocean, involve a variety of processes that may be analyzed and classified under different viewpoints, as described in the following.

A middle size watershed of temperate zones (but in fact applying to other climates as well) is schematically depicted in Fig. 1. To give an idea of what usually takes place at different elevations and distances along the watercourse, the longitudinal dimension is approximately indicated by a logarithmic scale, in such a way as to emphasize the complexity of the problems occurring at the smaller scales (farther and higher areas of the watershed).

Under the action of water (direct: rainfall, overland flow, channeled flow; and indirect: freezing and melting, infiltration, etc.), sediments are removed from the surface of the watershed and conveyed downstream. Depending upon the prevalent extension of the process in three, two or one spatial dimensions, sediment motion assumes three basic forms (mass, surface and linear), more or less corresponding, respectively, to (i) landslides, occasionally produced in the steepest slopes of the watershed, even if protected by vegetation; (ii) distributed soil erosion mainly occurring in undulated, scantily vegetated surfaces; and (iii) bedload and suspended transport by waterflow in the stream network. There are also a number of intermediate forms which share some characteristics with the basic ones, as for example: gully development (mass/surface/linear motion) rills erosion (surface/linear movement), debris flow (mass/linear motion). Wind is often the most effective cause of surface erosion where rainfall is extremely scarce, as in the desert or in arid zones.

Physical phenomena related to sediment motion are therefore extremely numerous and strictly connected with the morphoclimatic conditions under consideration. Moreover, they are traditionally dealt with by different disciplines and professions, very often under a quite “parochial” perspective.
Mass movement, characterized by quick and short displacements of large portions of soil, represent sometimes a risk for human settlements and infrastructures, but also a physiological source of sediments to the rivers in several natural watersheds (e.g. in alpine and humid tropical regions). Investigation on mass movement is generally carried out by applied geologists and, for the structural aspects, by soil mechanics engineers. Mass movement specialists are often barely interested in the final destination of the removed material as sediment yield.

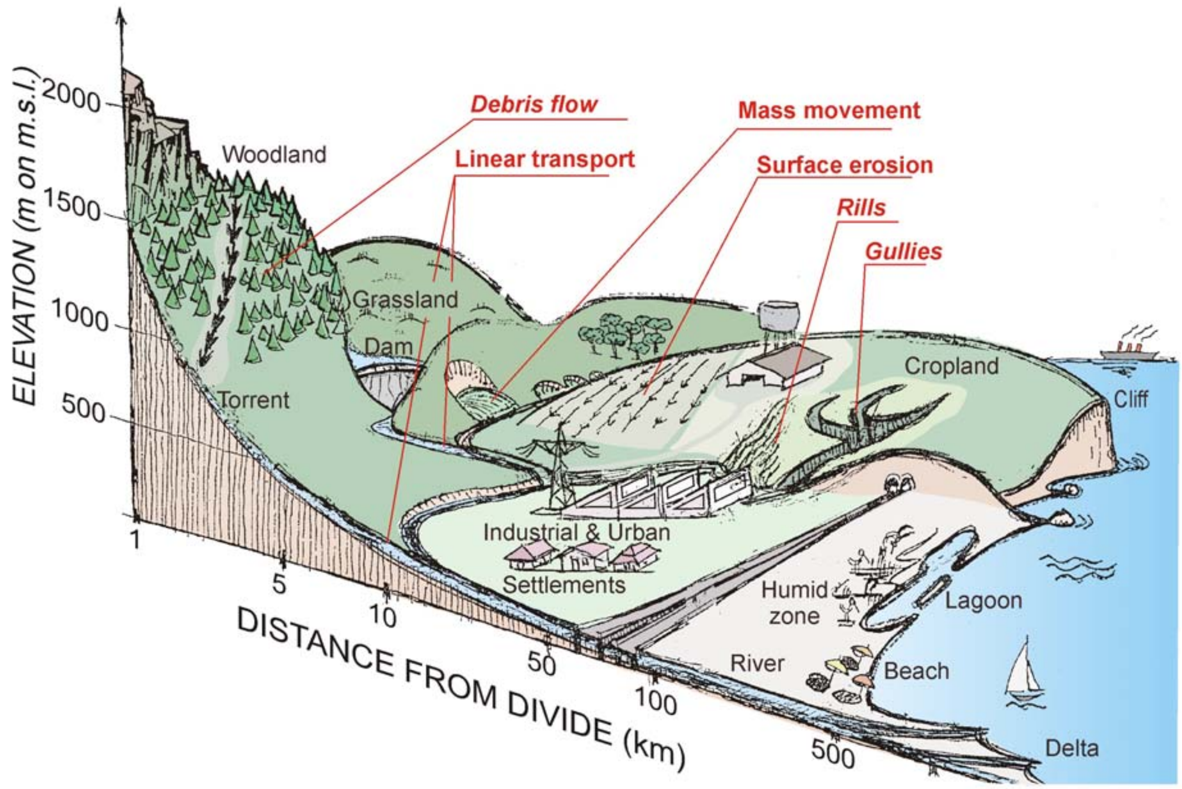

Fig 1-1: Sketch of a watershed in temperate zones: Basic forms of sediment motion

Surface erosion, usually as sheet erosion but including also intermediate forms like rill and gully erosion, because of its strict implications with land use and agricultural practices, usually belongs to the province of agronomists and agricultural engineers. It is also investigated by various scholars of earth science. These forms of erosion constitute a natural source of sediments both in arid tropical and temperate regions where rainfall is generally the dominant mechanism of sediment production. On the other hand, surface erosion tends sometimes to be overestimate as a component of sediment yield even in the cases where mass movement prevails.

Finally, linear transport is traditionally in the competences of hydrologists and river engineers. Bedload and suspended sediment transport convey coarse and fine particles over extremely long distances along the river, down to the estuary, the sea and the adjacent beaches, where they usually pass under the "jurisdiction" of coastal engineers and oceanographers. While solid transport in the river includes material produced by the entire watershed, fluvial and maritime specialists generally pay little attention to the sediment sources.

A specific disciplinary approach is almost invariably applied successfully to solve most of the engineering problems. However, to understand the behavior of sedimentary systems at relatively large space- and time scales (see Sect. 5), knowledge and experiences from different branches of science and professions should be brought together. This operation is not at all easy, not only between academic disciplines but also between separate ministries and agencies which do not have competence on sediments in every country.

A short review of the three basic forms of sediment motion mentioned above is given in the following Sections 2, 3 and 4.

2. MASS MOVEMENT

Mass movement corresponds to the detachment of sediments as a bulk from their original position (landslides), when the resisting forces (friction and cohesion) become lesser than the acting force (gravity). Mass movement is an important source of material for many rivers and in some cases, the most important one. In *humid tropical forests* as well as in *alpine climates*, for example, the natural thick vegetation cover is such that the direct effect of rainfall (kinetic energy) on the soil is negligible and the sediment production by surface erosion is practically zero. Yet the sediment transport by mountain rivers may be substantial and even extremely large (up to 10^4 t/km^2/year), due to the contribution of repeated *slope collapses* and occasional big *landslides*. Small and large mass movements from the watershed slopes typically occur during large floods and intense storms, and are often associated with *mud*- and *debris flows* in the upper branches of the hydrographic network.

Mud- and debris flows (including ash flow or "lahars", taking place along the steepest channels of volcanoes) are intermediate forms of sediment motion, between mass movement and linear transport, which require a relatively small minimum steepness to be initiated. While their motion depends on particle- and fluid dynamics (similar to linear transport), their triggering is controlled by static forces, basically depending upon friction, cohesion, slope and the degree of saturation of permeable material (as for mass movement). For this reason attempts have been made to model the triggering of both shallow landslides and debris flows by simulating the saturation process of the surface layers of watershed slopes and steep channels (see for instance Dietrich, W.E. and Montgomery, D.R., "SHALSTAB – a digital terrain model for mapping shallow landslide potential – to be published as a technical report by NCASI (National Council for Air and Stream Improvement). Also available on the Internet, Baum et. al, 2002.

The collapse of river banks is also controlled, like landslides, by friction, cohesion, slope and saturation, but its triggering is often determined by foot erosion produced by water flow. For this reason bank collapse material is considered part of linear transport (see Sect. 4) and its simulation usually included in morphodynamic modelling of rivers (see Sect. 6).

3. SURFACE EROSION

Surface erosion, prevalently developing over two dimensions, is definitely the most important source of sediment production wherever vegetation does not provide a sufficient cover of the soil from the rainfall impact, and morphological conditions are such as to foster the removal of particles by overland flow. This means that surface erosion is particularly active in cropland areas, especially where the type of soil is more vulnerable, yet erosion-control measures and correct cultivation practices have not been applied. In many temperate countries, an extremely high rate of surface erosion took place in historical times, following the rapid expansion of cultivated areas and before sustainable land management was adopted. The most recent episodes of this type occurred about hundred years ago in the United States, where extensive areas of the Midwest were rapidly transformed from natural grassland into cropland. For this reason soil erosion was first investigated at scientific and technical levels in this country, with special reference to the undulated landscape and climatic conditions typical for these areas.

The most active institution in this field was certainly the U.S. Department of Agriculture (USDA), where the renowned Universal Soil Loss Equation) (USLE). model was proposed. The USLE was developed several decades ago (Wishmeyer and Smith) by using the USDA database containing a very large number of results. The multiplication structure of the formula tries to put into account all the following factors: kinetic energy impact of rainfall combined with the intensity of rainfall, this last proportional to overland flow discharge (erosivity factor, R); resistance of the soil, quantified by means of descriptive tables (erodibility factor, K); slope length, also proportional to overland flow discharge (*length factor*, *L*); slope steepness, related to overland flow velocity (*steepness factor*, *S*);

protection by vegetation depending on plants, crop and vegetative phase (*cover factor*, *C*); and management practices (*practice factor*, *P*).

The USLE has been thoroughly criticized and defended in the literature, but also extensively applied even outside the USA, although very often with various adaptations. The formula provides, in principle, the values of sediment production at the "plot- or field scale" for a given period of time. For obtaining the corresponding data at catchment scale, the sediment production should be routed downhill to the hydrographic network and, eventually, downstream along the river to the closure section of the basin. The routing process that transforms the local *sediment production* into the integral *sediment yield* of the entire watershed is a rather delicate matter (see Sect. 8).

Besides the USLE equation, more sophisticated models such as ANSWERS, WEPP, SHE-SED, EUROSEM, etc. have been recently developed for simulating, at catchment level, the detachment of soil particles by rainfall and their subsequent transport by overland flow and by river flow over the entire catchment (Beasley et al. 1980, Nearing et al. 1984, Wicks and Bathurst, 1996, Morgan et al, 1996, etc.). In contrast with the so-called empirical models (like USLE), these models are usually called "physically-based" since they are constituted by theoretical differential equations (expressing the mass balance of water and sediments) and by appropriate algebraic equations (describing each of the physical processes involved).

Physically-based models resemble somehow the erosion-transport-deposition models employed in river morphodynamics (see Sect. 6). The physical processes involved in both *water flow* and *sediment motion*, however, are much more complicated on the watershed slopes than in rivers, and therefore much more difficult to be realistically simulated (see Sect. 6 and 8). For this reason, empirical models controlled by few overall coefficients (scarcely recognizable from the physical point of view but quite consistent and confirmed by many experiments) frequently give much better results than physically-based models controlled by a large number of coefficients (generally unknown and based on hardly plausible physical and geometrical schematizations) which ignore existing relevant interactions.

4. LINEAR TRANSPORT

Linear transport, namely taking place along one prevailing (longitudinal) direction, is the motion of sediments produced by persistent, channelized water flow. It is mainly responsible for river processes in the hydrographic network.

4.1 <u>Modes and rate of transport</u>

Linear transport assumes various modes (bedload, suspension and intermediate forms), but attempts have been made towards a conceptual unification of these forms, through the notion of adaptation length. The adaptation length expresses the distance required by clear water entering a uniform flow stream flowing over a uniform grain size bottom to reach the uniform sediment transport conditions. The adaptation length depends on the particle grain size and on the characteristics of the water flow, i.e. more precisely on the ratio between friction velocity u* and particle settling velocity w_s. When the ratio (u^*/w_s) is very small, the adaptation length has the order of magnitude of 10^2 grain diameters and the particles move by sliding and rolling as bedload. When this ratio increases, the adaptation length also increases correspondingly and the motion passes from saltation to suspension. Adaptation length is practically zero for coarse material moving as bedload, while for fine particles moving in suspension, it may reach the value of tens of kilometers.

The *solid discharge* of a natural stream (expressed by the mass or volume of sediments conveyed per unit time through a given cross section) may be somehow evaluated by the so-called *sediment transport formulas*. As it will be seen better in Section 7, most of the formulas have been obtained in laboratories under *uniform conditions (uniform transport by uniform plane flow* and *uniform grain* *size* material). In these conditions, the *total solid discharge* can be expressed as a function of the water flow characteristics and the particle diameter, but the total amount may be split somehow between bedload and suspended transport. In fact, the distance covered by the particles under the action of the water flow does have a statistical distribution, depending on the ratio (u^*/w_s). This ratio therefore defines the ratio between the number of particles instantaneously subject to different modes of transport, as well as their adaptation length.

When the adaptation length is quite long, the sediment transport rate does not depend solely on the local hydrodynamic and sedimentological characteristics, but also on the conditions upstream. This circumstance in part explains why the suspended transport in a given cross section of a river is often scarcely correlated with the local water flow.

The adaptation length can be evaluated by different approaches (Galappatti, 1985, Armanini and Di Silvio, 1988, Bolla Pittaluga and Seminara, 2003) and its effect should be taken into account, when necessary, in sediment transport computations (see Sects. 6 and 7).

4.2 <u>Sorted material</u>

In real rivers, particle grain sizes are more or less non-uniformly distributed, with markedly different statistical distributions for *bed material* and *transported material*. In general, bed material appears to be coarser than transported material, and the two distributions can be mutually related by considering the transport of each grain size class (see Sect. 7).

When treating different grain size classes, due attention should be paid to the interference of particles of different diameters. In sediment mixtures, in fact, the intrinsic larger mobility of finer particles is somewhat diminished by the protection of the coarser ones ("hiding" effect) while the intrinsic smaller mobility of coarser particles is augmented by their protrusion with respect to the smaller ones ("exposure" effect). With very strong water flow in flood periods, the hiding-and-exposure effect may even lead to an "almost equal mobility". In low flow periods, by contrast, the different intrinsic mobility of various diameters strongly prevails over the hiding-and-exposure effect (indeed, the coarser particles may even not move at all). In any case, over a long period of time, the transported material (e.g. the material intercepted by a reservoir) appears to be definitely finer than the average composition of the river bed.

The "hiding-and-exposure" effect may be taken into account by various empirical coefficients to be introduced into the formulas developed for uniform material. The time evolution of bed- and transport composition is usually modelled by resorting to the *active layer* concept, first proposed by Hirano and subsequently incorporated into many morphodynamic models. More sophisticated approaches have been developed more recently, either by disaggregating the bottom *active layer* into a *mixing-* and an *intrusion layer* (Di Silvio, 1991), or by considering the bottom a continuous, indefinitely deep layer, statistically described in terms of entrainment capacity (Armanini, 1995, Parker et al, 2000)

4.3 <u>Cohesive material</u>

In some circumstances (e.g. estuaries, flood plains, deep reservoirs) sediments can hardly be considered non-cohesive. The role of cohesion is quite important both in the deposition phase (flocculation) and in the re-entrainment process (compaction). The pioneering work of Partheniades and Krones in the 1960s and of Methas and Partheniades in the 1970s is still the foundation of many models for cohesive materials. Some of the models developed from these basic concepts, however, do not appear to be completely satisfactory and are unable to explain a number of phenomena observed in nature. There is therefore much interest in the attempt by Winterverp (2001) to bring together the behavior of cohesive and non-cohesive material within a unified physical framework with specific definitions of vertical fluxes for each type of sediment.

5. TIME AND SPACE SCALES OF SEDIMENTARY SYSTEMS

Morphological processes may be seen as the product of the repeated succession of three phases of sediment motion: erosion, transport and deposition. In some cases, one of these three phases is definitely dominant. For example, soil removed from short watershed slopes, either by surface erosion or mass movement, may never be replaced by other soil. Conversely, sediment trapped by a deep lake or sea is not entrained and put into motion anymore. In these cases, the erosion or deposition process is time-depending but monotone (namely producing either a progressive degradation or a progressive aggradation). In many other cases, by contrast, subsequent phases of erosion, transport and deposition take place sequentially on the same location, giving origin to complicated alternating morphological processes. In this last case one can speak only of *net* degradation or aggradation of a certain *sedimentary system* over a prescribed *period of time*.

When considering morphological processes, it is important to have in mind the time- and space scales under consideration. The repeated succession of erosion, transport and deposition may concern for example: (i) the sliding, rolling and saltation of sediment particles over bed ripples (space scale: boundary layer thickness, say millimeters); (ii) the propagation of dunes (space-scale: river depth, say meters); (iii) the formation of bars and meanders (space scale: river width, say hundreds of meters); (iv) the general aggradation or degradation of a river (space scale: watershed, say up to thousand kilometers). The time-scale of each system may be associated to the corresponding space-scale, *via* a typical process velocity.

It is important to note, in any case, that each system at a given scale may be considered a component (or sub-system) of the system at the larger scale. The morphodynamics of a component does in principle interact with the morphodynamics of the systems at both larger and smaller scales. However, to describe the behavior of a component (e.g. the propagation of dunes along a river reach) it is usually assumed that the system at the larger scale (e.g. bars and meanders) remains *stationary* at the time-scale of interest for the component (dunes). At this time-scale, conversely, one assumes that the subsystem at an even smaller scale (e.g. bed ripples), although non-stationary, is in an *equilibrium condition* with the larger system (dunes). This simply means that, during the propagation of the dune, single ripples may appear or disappear, but their statistical distribution (and consequent hydraulic roughness of the dune surface) depends exclusively on the dune configuration. This assumption is only valid, in principle, when the relevant systems and sub-systems have markedly different scales, yet it is implicitly assumed in most morphological models (see Sect. 6).

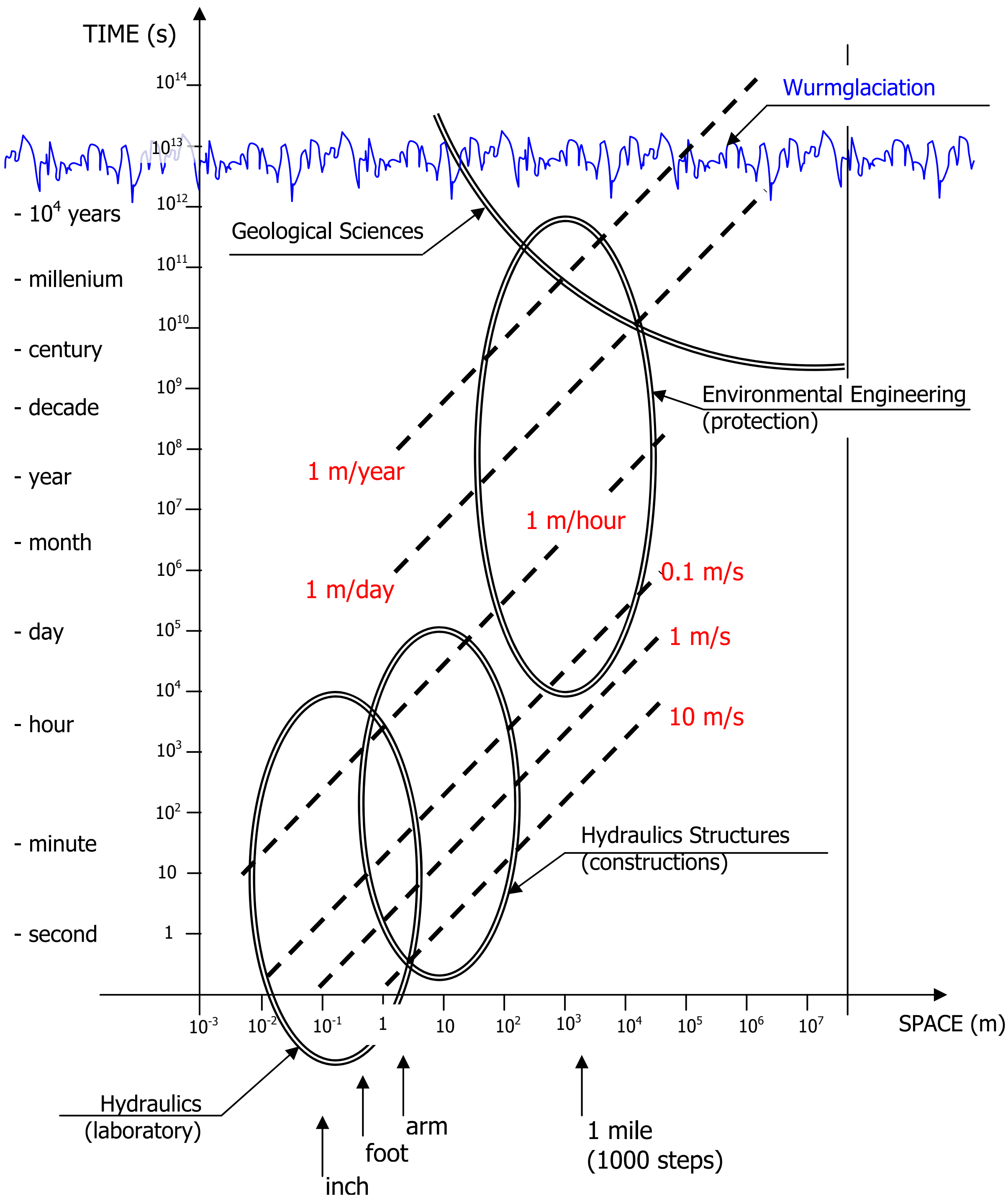

Fig 5-1: Time- and space scale of sedimentary systems

The scales of morphological processes extend over several orders of magnitudes, ranging from microns to continental sizes (in space) and from seconds to millions of years (in time). The graph in Fig. 2 indicates the range of interest for various disciplines interested in sedimentary systems. For *Hydraulic Structures* (*construction prototype)* engineers are generally interested in problems defined (in space) by the "size of the structure" and (in time) by the "event duration", or, at most, by the "project life" of the structure. For *Hydraulic Laboratories (laboratory experiments)* the range of interest is defined by the facility's size and the process velocity. In basic research (e.g. for analyzing the behavior of individual sediment particles) the relevant sizes may be extremely small, while for physical models they are generally larger, although obviously much smaller than the size of the corresponding prototype structure (we may say, in Froude similitude, 100 times less in space and 10 times less in time). However, if the engineer wants to assess the morphological effects of the structure he has designed on the entire river system, he should take into account much larger scales. For example, the presence of the Aswan Dam is already perceived, after several decades since its construction, in the Nile's Delta (subject to erosion) which is thousands of kilometers downstream. Yet, the adaptation process of the entire Nile River system will take an extremely long time to attain a new quasi-equilibrium configuration. In other words, as shown in Fig. 2, the time- and space scales for *Environmental Engineering (protection)* tend to be much larger than for hydraulic structures and be closer to those of the *Geological Sciences*, namely "geological times" and "continental sizes".

As a pure indication, the time axis of the figure is bounded by the end of Würm glaciation, the last large climatic change in Europe before the present changes which has impacted both hemispheres, with some local variation; and in terms of sea level change, the entire planet. We may assume, in fact, that at the geological scale the climatic forcing after the Würm glaciation was reasonably stationary. By contrast, if we consider a longer period of time, several processes would appear to be controlled by non-stationary climatic conditions (sequence of glaciations and consequent sea level changes), as well as by variable phases of tectonic uplift and subsidence.

However, even by limiting the analysis to the last ten thousand years, a quite large number of time- and space scales controlling the behavior of sedimentary systems should be considered when developing morphological models (see Sect. 6).

6. MORPHOLOGICAL MODELS

A large number of morphological models developed at different time- and space scales, and with various degrees of detail and approximation are available in the literature. In this section attention will be concentrated on the modelling of linear transport phenomena in particular (see Sect. 4). Models of mass movement and surface erosion are briefly mentioned in Sections 2 and 3.

6.1 Small scale models

Detailed small scale models have been developed specifically for research purposes. Many of these models (still in their infancy and needing to improve) have the scope to reproduce the movement of individual particles under the action of other particles and water flow and are usually based on a Lagrangian approach. They should be able, in principle, to reproduce the behavior of small scale systems (microforms up to the river depth-scale) and may be extremely useful to explain the hydraulic resistance mechanisms (grain and form roughness), to show the validity and limitations of transport formulae, to investigate the dynamics of movable bottom and to describe the motion of hyper-concentrated liquid-solid mixtures.

6.2 Intermediate scale models

These models are most commonly used for practical applications. They are typically extended to the size of a river reach and are applied for relatively short time durations (from a single flood event to a few years). As mentioned in Section 5, during this time all the processes at subsystem scale (microforms, hydraulic resistance, sediment transport rates, etc.) are incorporated *via* simple predictors, usually "equilibrium" algebraic equations, as a function of the water discharge. Conversely, all the processes at larger scale (climate, watershed configuration, etc.) are supposed to be stationary.

Intermediate scale morphological models are obtained by averaging convection-diffusion equations for sediments and the Reynolds equations for water (in their turn obtained by averaging the water continuity and Navier-Stokes equations over turbulence) over appropriate space dimensions. The most common commercial models are *1-D* (*one-dimensional)*, i.e. averaged over the river cross-section (but possibly disaggregated into a number of sub-sections). One-dimensional models can simulate bottom erosion and deposition along the river (generally the most relevant requested information), somehow "re-distributed" over the cross-section. One-dimensional models can be easily applied to relatively large portions of the hydrographic network or even to rivers and tributaries of an entire watershed.

Nowadays, however, *2-D (two-dimensional)* models are becoming rather common, i.e. averaged over the river depth, also available as commercial codes developed by several laboratories. Two-dimensional models can in principle simulate all the process at the width-scale (migrating and stationary bars, braiding and bifurcations, sediment exchange with flood plains, etc.). Bank collapse and reconstruction can also be incorporated into a 2-D model, which therefore will be able to reproduce meander formation and propagation.

Secondary currents over the cross section are important for localized scouring (piles, groynes, etc.) and in meander morphodynamics. Their reproduction requires a *3-D* (*three-dimensional)* model in principle but, at bar/meander scale, their local effect can be approximately accounted for by a 2-D model. Reproduction of density currents, often important in certain reservoirs, also requires a 3-D model. In some cases, however, a *vertical 2-D* model (i.e. averaged over the reservoir width) can also be considered.

Intermediate scale models, either 1-D, 2D or 3-D, are extremely sensitive to the *boundary conditions* to be prescribed at the upstream and downstream ends of the river reach under investigation. Correct boundary conditions for morphological models (de Vries, 1993) should be given in terms of sediment input of each grain size fraction (at the upstream end) and in terms of either water-level or bottom-elevation, respectively for sub-critical and super-critical water flows (at the downstream end). Note that boundary conditions depend in principle on what is going on upstream and downstream respectively of the considered reach. For relatively short simulations (years), sediment input upstream can be evaluated by a reasonable hypothesis based on "local" quasi-equilibrium conditions (see Sect. 7); the same can be made for water level or bottom elevation downstream. For longer simulations (centuries), however, the behavior of the entire river system should be accounted for explicitly (Sect. 8).

6.3 <u>Large scale models</u>

Although 1-D models have sometimes been applied to relatively large real watersheds for specific flood events, not many examples are available in the literature of morphodynamic modelling at very long (historical or geological) time-scales, except in a few very schematized situations (simple geometry, constant waterflow, uniform grain size). The effects of geometrical, hydrological and sedimentological *non-uniformities,* invariably present in real systems, have been investigated only in part for long-term, large scale simulations of actual rivers and relevant watersheds. In fact, averaging "non-uniformities" of any type in non-linear equations produces "residual terms" which should be properly assessed and eventually modelled with appropriate sub-models (Di Silvio and Marin, 1996).

It may be of interest, in this respect, to explore the possibilities offered by *long-term, large scale morphological models* where averaging is performed on time (year or number of years) and/or space (river reaches of various length). In practice, these models filter the shorter morphological fluctuations and compute only the long-term, large scale evolution. Long-term models have been especially developed for estuaries, but could in principle also be applied to river systems.

7. SEDIMENT TRANSPORT FORMULAS

A fundamental component of any morphological mathematical model is the predictor of the sediment transport rate as a function of sediment grain size and flow characteristics. Sediment movement produced by channelized water flow (the so-called *linear transport*, introduced in Section 4), may be modelled in detail, from the *particle scale* up to the *depth scale* (ripples, dunes, hiding-and-exposure, etc.), as mentioned in Section 6.1. In practice, what can be utilized for *intermediate scale* modelling

(Sections 6.2 and 6.3) is a variety of algebraic equations (the so-called “transport formulas”), sometimes associated with other formulas providing the size of bedforms and the corresponding hydraulic resistance.

The available transport formulas have been obtained, above all, from laboratory flume experiments carried out in uniform conditions (uniform flow and transport and uniform grain size). Since the early work of Du Boys (1879), a large number of transport formulas have been proposed by several authors. Since all these formulas have been obtained in specific experimental ranges of flow and sediment characteristics, it is no wonder that they appear inaccurate when applied to other situations.

Assessing the prediction capability of different formulas, or even recognizing the validity limits of each one, is not an easy task. At first glance all the formulae seem to be hardly comparable. Yet it may be interesting to perform a dimensional analysis of their structure, assuming that the process is controlled by 6 *independent* variables (e.g. shear stress u_*, grain size d, density of grain ρ_s, density of fluid ρ, viscosity ν and gravity g).

Let us define (see for example Yalin, 1977) the non-dimensional sediment transport rate:

$$T_* = \frac{q_s}{d\sqrt{g\Delta d}} \tag{1}$$

where q_s is the solid discharge in volume per unit width and $\Delta = \frac{(\rho_s - \rho)}{\rho}$ is the relative density of sediments. It follows that T_* should be a function of 3 independent non-dimensional morphological parameters; for example:

- the particle Froude number (or mobility index, or Shields parameter):

$$F_* = \frac{u_*^2}{g\Delta d} \tag{2}$$

- the particle Reynolds number :

$$\mathrm{Re}_* = \frac{u_* d}{\upsilon} \tag{3}$$

- the relative depth :

$$\left(\frac{h}{d}\right) \tag{4}$$

The particle’s Reynolds number plays an important role for fine particles transported in suspension, as it controls the settling process. The relative depth, by contrast, is important for very coarse particles moving as bed load, since they can affect the free surface of shallow flows. Conversely, as for several hydraulic phenomena, the influence of both the Reynolds number and the relative depth tends to disappear when these quantities become very large. This occurs for mountain rivers (high flow velocity and coarse material) and for large plain rivers (large depth and fine material) respectively.

In any case, the parameter F_* is invariably the most important one as it represents the ratio between the mobilizing effect of the water drag on the particle and the stabilizing effect of the particle's immersed weight. Most of the available transport formulae, in fact, can be approximately plotted on a graph T_* vs. F_* , either in the form

$$T_* = aF_*^{\beta} \tag{5}$$

or in the form :

$$T_* = b\left(F_* - F_{*cr}\right)^{\gamma} \tag{6}$$

assuming that the coefficients *a* and *b* and the exponents β and γ are not constant, but functions of other quantities besides F_* , as in fact particle Reynolds number and relative depth.

The *monomial structure* of eq. (5) is typical of many formulae like those of Kalinske (1947), Brown (1950), Engelund and Hansen (1966), etc. The *binomial structure* of eq. 6 implies that no movement occurs if the mobility (or Shield's) parameter F_* is smaller than a critical value F_{*cr} (in principle, function of Re_* and h/d). The binomial structure is typical of several popular formulae like those of Meyer-Peter and Mueller (1948), Ackers and White (1973), van Rijn (1975), etc. Note also that other formulae in literature, having an apparently different theoretical background (like the ones based on minimal stream power, for instance Chang, (1977), Yang, (1976), can approximately be written as eqs. 5 and 6.

It is a matter of philosophical discussion whether in principle a critical value F_{*cr} for incipient sediment transport should exist at all. Indeed, due to the stochastic character of turbulence, one may think that an (occasional) transport would even take place with extremely small (average) values of F_*. In any case, since the transport rate should rapidly decrease for very low values of F_*, the exponent β in monomial formulae needs to be rather large (and in fact ranges between 2.5 and 3), while the exponent γ in binomial formulae is much smaller (ranges between 1.2 and 1.5).

For both types of formulae, however, the numerical value of the exponents (β and γ) and of the coefficients (*a* and *b*) should depend, explicitly or implicitly, on the other non-dimensional parameters mentioned before, that is *Re** and (h/d). In fact, many of the experimental formulae contain other quantities besides F_* that affect the non dimensional sediment transport. Although these quantities are not explicit functions of *Re** and (h/d), they are very likely related somehow, depending upon the range of flow- and sediment characteristics in which the experiments have been carried out.

At this point, for practical applications, instead of selecting a certain available formula, it is perhaps better to resort to an expression like (5) or (6). In this case, of course, the values of exponents and coefficients should be properly chosen for the river configuration one is interested in (ranging from steep alpine torrents conveying gravel and boulders, to slow lowland rivers conveying silt and sand). This choice is in fact a sort of transport formula calibration.

For calibrating the transport formula of a given river configuration, the simpler monomial equation (5) is preferable to the binomial equation (6), even if one has to expect a larger variability for the values of *a* and β than for the values of *b* and γ. The calibration procedure of the transport formula (Di Silvio, 1996) consists in associating to eq. (5) a uniform flow formula, either Chézy or Manning-Gauckler-Strickler, and in introducing the grain size distribution of the bed material together with an appropriate "hiding-and-exposure" coefficient (Sect. 4.2). The hiding and exposure coefficient, multiplying the value of q_{si} (solid discharge of the i-th grain size) in eq. (1), is assumed here as

$(d_i/d_m)^s$, where $d_m = \sum \beta_i d_i$ represents the mean grain size of the bottom material, with $0 < s < q$ an appropriate exponent and β_i the percentage of the i-th grain size class present in the bottom. With respect to the uniform grain size material, the "hiding-and-exposure" coefficient slightly augments the "intrinsic mobility" of coarse particles ($d_i > d_m$) and diminishes that of the finer ones ($d_i > d_m$).

The final expression for the total sediment discharge (sums of all grain size classes, $i=1,2,\ldots N$) is :

$$Q_s = \sum Q_{si} = \alpha \frac{\sum \beta_i d_i^{\,s-q}}{(\sum \beta_i d_i)^s} \cdot \left(\frac{Q^m I^n}{b^p} \right) \tag{7}$$

where Q, I and b are respectively the waterflow discharge, the energy slope and the river width, d_i is the diameter of the *i*-th grain size class and β_i is the percentage of the *i*-th grain size class present in the bottom. The coefficient α incorporates all the quantities assumed as "constant" in the above mentioned procedure. The value of the exponents m, n, p and q depends on the exponent β in eq. (5) and on the selected uniform flow formula, according to the following expressions:

Chézy

$$\begin{aligned} m &= \frac{2}{3}\beta \\ n &= m \\ p &= m-1 \\ q &= \frac{3}{2}(m-1) \end{aligned} \tag{8}$$

Manning-Gauckler-Strickler

$$\begin{aligned} m &= \frac{3}{5}\beta \\ n &= \frac{35}{30}m \\ p &= m-1 \\ q &= \frac{3}{2}(m-1) \end{aligned} \tag{9}$$

In general, assuming a constant Chézy coefficient (i.e. a constant *relative* roughness) is more appropriate for lowland rivers (dominant dune resistance); while a constant Manning coefficient (i.e. a constant *absolute* roughness) is more appropriate for mountain rivers (flat bed and dominant grain resistance). The exponent s of the hiding-and-exposure coefficients tends to increase for strongly sorted material (mountain rivers) and to become equal to q for extremely high values of Q (equal mobility). It may be taken equal $s = 0.8$ in torrents and much less (down to almost zero) in many plain rivers.

Note that eq. (7) is just another form of eq. (5), in which the transport of each grain size class present in the bottom has been considered. Equation (7) indicates that, the other quantities being constant, an univocal relation should exist between Q_s and Q. This is true, however, only for an experimental flume in *equilibrium conditions* (uniform flow for water and sediments): indeed, for a re-circulating flume with prescribed values of I, b and bottom composition, the transport rate Q_s is a unique function of the water flow Q.

For a real river, by contrast, different values of Q_s are measured for the same value of Q. This is basically due to the fact that the local energy slope I and the local bottom composition β_i may vary during the hydrological cycle, as fluctuating erosions and depositions invariably occur. As seen before, moreover, the exponent s may also not be constant. Finally, if the material is very fine, the material transported in suspension may be not solely controlled by the local conditions, but also by the conditions upstream (see Sect. 4.2). The last circumstance, however, is not so dramatic if the "adaptation length" is shorter than the river reach under investigation.

Eq. (7) indicates that the composition of transported material (*Qsi/Qs)* is much finer than the local bottom composition (β_i). This means that the total transport formula gives reason for a relevant transport of very fine particles, even if their presence in the bottom is extremely scarce. As it appears from eq. (7), in fact, due to their much larger mobility, the particles belonging to a very fine faction (say d_i = 50 microns) may have a very small value of β_i, but a very large value of $\left(Q_{si}/\sum Q_{si}\right)$. In other words, the following accepted notion may be misleading: that only the transport of the material abundantly represented in the river bed (the relatively coarse, so-called “bed material”) depends on the local conditions, while the fine material should be considered “wash-load”. By contrast, even the so-called “wash-load” leaves a trace in the bottom composition that can be used to compute the total transport.

In conclusion, for relatively large watersheds, the scattering of short-term measurement Q_s vs. Q is generally due to short-term fluctuation of I, β; and (probably) exponent s, rather than to the time-dependent input of fine sediments from the watershed slopes.

Indeed, if one supposes that fluctuations of the above mentioned quantities are mutually independent and assumes an exponential duration curve for *Q(t)*, the integration of (7) over one year provides:

$$V_s = \frac{\alpha}{m}\frac{I^n}{b^p}\frac{\sum \beta_i d_i^{\,s-q}}{(\sum \beta_i d_i)^s}\cdot\left[Q_o^{m-1}V_o\right] \qquad (10)$$

where V_s is the total annual transport of sediments (all classes), V_o is the annual runoff volume and Q_o the annual flood peak.

Although the hypothesis on the statistical independence of the fluctuations of I and β may be questionable, experimental applications to real measurements show a very good correlation between hydrological parameters Q_o and V_o and the annual sediment yield V_s, with an exponent m between 1.5 and 2.5 (depending on the type of river). The structure of eq. (10) is particularly convenient for calibration against sedimentation data in reservoirs.

8. SEDIMENT YIELD AND SEDIMENT PRODUCTION

One of the most difficult problems in establishing a sediment balance at watershed scale is the relationship between the sediment removed from the watershed slopes (soil production) and the soil transported by the river (sediment yield). The very same definition of those quantities may in fact present some ambiguities.

A possible unambiguous definition of *sediment yield is the total amount (mass or volume) of sediments, of any size and origin, transported as bedload or in suspension through a given cross-section during a certain period of time* (year, day, flood event, etc.). Very often, however, sediment transport is disaggregated into two parts: the so-called “bed-material transport” (typically coarser than a conventional grain-size limit, say between 20 and 80 microns) and the so-called “washload” (below that limit). While the transport of bed-material is supposed to be a function of riverbed composition and flow characteristics, washload is assumed to be fed into the river from the watershed slopes and conveyed downstream by the river flow, with the same velocity as that of the water, i.e. without any interaction with the bottom. In many instances, washload (defined in this way) turns out to be a very large portion of the total transport, so that “sediment yield” is practically equivalent to the corresponding “sediment production” during the same period of time.

The distinction between bedload material and washload is obviously made for the sake of simplification but, as mentioned in the previous section 7, it does not have a solid physical foundation. Indeed, even the finer particles have multiple phases of transport, deposition and resuspension and their average motion is much slower than the water’s. Consequently the sediment yield of the river may be much less or much larger than the sediment production during the same period of time.

Let us now consider the definition of sediment production. On the analogy of sediment yield, a straightforward definition of *sediment production is the total amount of sediments, of any size and origin, detached by surface erosion and mass movement, from a given location of watershed and transported downhill during a certain period of time (year, month, storm event, etc.).* In this way, sediment production is expressed as entrainment per unit surface but, in practice, the amount of removed particles can only be measured, as transport per unit width of the watershed slope, at a finite distance more or less remote from the closest “divide”. It is apparent in fact that a “punctual measurement" of sediment production independent from the transport does not have much sense and that some space-averaging operation should be performed over the slope surface. Experimental data, indeed, are never available point by point, but at “plot” or “field” scale (for cropland) or at “slope” scale (for natural watersheds).

As already observed, however, space-averaging is not at all a banal operation. First of all, even for extremely tiny pieces of slope surface, detachment and transport processes are independent but inevitably related. At a *small scale* (overland flow depth) we may observe that a thin overland flow cannot maintain a stable fully two-dimensional aspect but invariably tends to concentrate into a channelized flow which collects and transports the material detached by the raindrops from the closest surface. This is very apparent for rills and gullies, but even "diffused" sheet erosion actually occurs though embryonic and intermittent micronetworks, basically controlled by vegetation. For larger and larger sizes, as it conveys larger and larger concentrated waterflow, the micronetwork tends to become more stable and well defined and to evolve towards the permanent, morphologically controlled hydrographic network.

At an *intermediate scale* (experimental plot, field or natural slope) both the runoff and the sediment transport, actually concentrated along the micronetwork, are somehow integrated (i.e. averaged) over the relevant surface, as they are generally measured at the surface foot. A complete and reliable set of such data has been formed, over decades and decades, by agricultural engineers on experimental plots in many countries of the world with different soils and different crops. Experimental plots have in general a narrow rectangular surface with no transversal elevation gradient and a uniform longitudinal steepness. These data have been employed in the United States to develop the celebrated Universal Soil Loss Equation (USLE) and elsewhere around the world to adapt this formula to different agricultural and climatic conditions.

As discussed in Section 3, the USLE estimates the sediment production, in mass per unit surface, as the product of six factors which include the length of the plot L and the steepness S. While for an *experimental plot* or even for a regular *cropland field*, ditches clearly show where they initiate and terminate, for a *natural slope* the only apparent boundary is represented by the channels and the divides of the hydrographic network. For natural conditions it is therefore more practical, to define *the sediment production in a given (preferably small) hydrographic watershed as the portion of sediments, of any size and origin, detached by surface erosion and mass movement, which reaches the (permanent) hydrographic network during a certain period of time (year, month, storm event, etc.).*Sediment production of the watershed can be computed by applying the same formulae (e.g. USLE) calibrated at field scale from experimental plot data, where one assumes for the *length* and *steepness* of the natural slopes respectively the inverse of the basin's *drainage density* and *relief*. In this computation a certain reduction coefficient (*slope delivery ratio*) should be applied for taking into account the trapping effect along the natural slope, especially when the slope is quite long and its profile is undulated.

To transform *sediment production* into *sediment yield*, it would now be necessary to route the input of sediment all along the hydrographic network, down to the closure section of the watershed. With the previous definition, a distinction has been made between the intermediate scale (field or slope length) where sediment production takes place and the large scale (watershed or river length) where river processes take place. An even more aggregate definition of *sediment production is the portion of sediments … which reaches the closure of the watershed.* In this case, the computation at river scale should be affected by an even smaller reduction coefficient (*overall delivery ratio*), which should also take into account the river processes along the entire hydrographic network.

The concept of “overall delivery ratio” for sediments is somehow analogous to the concept of “runoff coefficient” for water. Yet it is much more difficult to be defined and predicted, due to its variability in space and time along the sediment route. In fact the very notion of overall delivery ratio is not much utilized in the recent literature. From the early data (Gottschalk and Brun, Shumm, Mauer, Roehl, Williams and Berndt,) it appears that delivery ratio decreases from one to a few percents, more or less proportionally to the inverse of the stream length (or square root of the watershed area) but scattering of data appears to be extremely high. Several attempts to obtain a more accurate prediction of delivery ratio as a function of the watershed and river morphology have been made (see for instance Walling and Webb, 1996) but did not generally provide valid results.

Similarly to the “runoff coefficient”, the concept of “delivery ratio” is hardly useful when it becomes much smaller than one (namely for watersheds larger than 50-100 km^2). The notion of delivery ratio is in fact probably acceptable exclusively at intermediate scale, namely for an overall description of the “monotone” trapping effect the watershed slopes, where very few localized permanent 'sinks' can only give rise to (averaged) values of the delivery ratio very close to one.

When river processes become dominant it would probably be better to substitute the static concept of "delivery ratio" by a dynamic concept of "response delay", in which the time scale also plays a role (Di Silvio and Marion, 1997). Indeed, if the watershed is large, it is not correct to assume that the very same particles detached from the watershed slopes during a certain storm can reach the closure section of the basin during the corresponding flood. The sediments moving as bedload or as suspended transport along the river (including the very fine ones usually called “washload”), have continuous phases or deposition and re-entrainment with the river bed, banks and floodplains. Repeated deposition and re-entrainment may produce relevant granulometric, altimetric and planimetric changes at different time scale and, in any case, will strongly delay the response of river morphology (and river transport) with respect to the sediment input from the watershed slopes.

An evaluation of sediment yield directly in the river can be made by utilizing regular (daily) measurements of turbidity and water discharge carried on at some stations. This procedure assumes that there is a direct relationship between “turbidity” usually measured in one single point of the cross section and "transport concentrations" (ratio between total sediment transport and water discharge). This hypothesis is probably acceptable, especially in the long term, but it deserves further theoretical and experimental consideration (Walling and Webb, 1988). Turbidity and water discharge measurements are often spatially scanty or incomplete in time. In some cases the data set may be integrated or extended by calibrating "river specific" transport formulas (see Section 7).

The most precious and reliable information about sediment yield in terms of both quantity and grain size composition, however, is given by the progressive sedimentation of existing reservoirs. The surveying technology based on the joint use of remote sensing and Global Positioning System (GPS) has already been applied in similar circumstances. In assessing the sediment volume trapped in a reservoir, the time-dependent compaction of the deposited material should be taken into consideration (see for instance Morris and Fan, 1998). The data collected in existing reservoirs, as well as at measuring stations, may also be used for calibrating reliable semi-empirical relationships (even if limited to a specific river configuration) which provide long-term sediment transport as a function of hydrological, geometrical and sedimentological characteristics of the river reach (see Sect. 7).

9. THE GEST (GLOBAL EVALUATION OF EROSION AND SEDIMENT TRANSPORT PROCESSES) PROJECT

One of the main purposes of the GEST project is the evaluation of the global sediment yield, to be evaluated in a significant number of cross-sections of the main rivers of the world. Besides the present conditions, the assessment should also be repeated considering past and future scenarios.

As discussed in the previous Sect. 8, the evaluation of sediment yield can be carried out, in principle, either directly (by measuring or predicting the sediment transport rate in a given section, based on the local characteristics of the river), or indirectly, by measuring or predicting the soil production in the watershed slopes and routing this sediment down to the closure section. The application of either method depends on circumstance, as it appears from the following limit cases. For very small and steep watersheds (e.g. a gully) and extremely fine materials (clay), the response of the system is very fast. In this case, the sediment transport in the closure section (wash load) practically coincides with the soil production. By contrast, for large watersheds and relatively coarse material, the sediment yield (even transported in suspension) is quite independent from the soil production during the same event but solely depends on the local characteristics of the river. For intermediate conditions the variable sediment production and the more or less delayed response of the river system should both be taken into account. While a single (constant) “delivery ratio” appears to be inadequate for a satisfactory reproduction of this process, a relatively simple modelling of the river network may suffice.

The GEST project represents a unique occasion to assess and compare different methods. By utilizing experimental data on both the river and the watershed, collected in a variety of morphological, hydrological and sedimentological configuration, an acceptable (general) methodology might be developed, as one of the results of GEST, for predicting sediment yield at global scale where direct information is not available.

10. SUMMARY AND CONCLUSIONS

In the framework of activities carried out by the International Sedimentation Initiative (ISI) Task Force Group, a brief review has been made of the state-of-the-art of knowledge of the dynamics of sediment erosion and sedimentation, focusing on the following aspects:

- Surface erosion, mass movement and linear transport: these three basic forms of sedimentary processes were examined by using a variety of different approaches, depending on the climatic, social and disciplinary environments, which is not usually the case.
- Sedimentary systems and sub-systems: review of how to cope with different time- and space scales and how to select the necessary morphological models with different degrees of detail (1-D, 2-D and 3-D models).
- Increasing importance of large scale, long term processes: review of how to incorporate these processes into engineering practice.
- Sediment yield and sediment production: review of how the "transfer function" can be (reasonably) modelled.
- The GEST project: review of sediment yield assessment at global scale which provides an occasion for testing assumptions and methodologies.

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A SOUTHERN PERSPECTIVE: EROSION AND SEDIMENT DYNAMICS FROM CATCHMENT TO COAST

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1. INTRODUCTION

At the ISI Steering Committee meeting in March 2004 in Paris, France, and in follow up discussions, it was decided to prepare short review documents discussing the state-of-the-art of research related to the erosion of sediment in the catchment and the transport through a river system to the coast.

The work was distributed between ISI representatives in China, Italy and South Africa, with the latter written up in this review document, which gives a “southern” perspective (versus a “northern” perspective). The idea was not to separate different ways of dealing with the sediment and sediment dynamics, but rather to address specific climatic conditions and sediment characteristics in various parts of the world which are quite different from developed countries where many of the hydraulic and sediment balance theories have been developed. In this sense “southern” perspective means to include and discuss aspects that are typically found in Africa and Australia, such as:

- Semi-arid climatic conditions
- Sediment yields consisting mainly of fine sediments where availability from the catchment is dominant, and not only sediment transport capacity plays a role
- Large catchments with limited data

This review of research initiatives should be seen as a first attempt to summarize many years of research carried out all over the world on this subject. Accordingly, methodologies have been briefly reviewed to describe the sediment transport related processes from the origin in the catchment, through the river and reservoirs, the estuary (or delta), down to the ocean.

2. WHAT ARE THE MAJOR CONCERNS REGARDING SEDIMENT YIELD IN SEMI-ARID REGIONS?

In semi-arid regions soil erosion is a serious concern that poses a threat to the sustainability of small scale and subsistence agricultural production through the erosion and washing away of the topsoil which leads to the loss of crop production media. Of particular significance is that rural communities which practise subsistence agriculture and communal grazing are most affected by very severe soil erosion. Land degradation and soil erosion also lead to accelerated storage losses due to reservoir sedimentation.

An increase of suspended solids concentrations in flowing water also causes degradation of the environmental quality of rivers. Depending upon their chemical composition, sediments may carry plant-usable nutrients such as phosphorus and other fertiliser residues from agricultural lands. Nutrient rich water leads to eutrophication in reservoirs and lakes. Eutrophication may, furthermore, lead to increased evaporation and hence water losses. The dense vegetation in reservoirs may also clog outlets of dams and kill aquatic fauna through reduction of dissolved oxygen. Sediments, particularly those which are derived from densely populated areas without proper sanitary facilities, may also carry pathogens such as *Escherichia coli (E. coli)*. High concentrations of suspended solids, nutrients and pathogens in water create the need for expensive purification, especially to make the water suitable for domestic and industrial (manufacturing) uses.

3. SOIL EROSION AND SEDIMENT YIELD DETERMINATION

In this section techniques on how to quantify soil erosion and sediment yield are described, based first on measured data (rivers and reservoirs), then as regional and empirical approaches, and in the end by physically-based watershed models.

3.1 River sediment load / discharge rating curves and cumulative plots

In regions where the sediment transported in the river is relatively coarse consisting of sand, gravel or coarser particles, it is possible to hydraulically determine the sediment yield. By using the local hydraulic conditions at a site such as the flow area, water depth and energy slope, it is possible to calculate theoretically the sediment transport capacity at a specific time. During some flow conditions when the reach upstream is a depositional reach, the transport capacity at the site can only be achieved by bed re-entrainment which can only happen if the critical condition for re-entrainment is exceeded. Thus, in a quasi-equilibrium river with coarse sediment, the sediment transport capacity and the observed sediment transport should be more or less in agreement. Yet, when finer sediment (clay and silt) also forms part of the transported sediment, the relationship between sediment transport capacity and actual sediment transport is poor. This is due to the fact that the fine sediment is availability-limited, especially in semi-arid catchments. The sediment particles are so small that the transport capacity of the stream far outweighs the sediment availability. Figure 3-1 shows the sediment concentration-discharge relationship at a site in South Africa. Only after prolonged droughts does fine sediment build up in the catchment again to increase the supply. Owing to availability, sediment loads are as a rule much higher at the beginning than at the end of the wet season. In semi-arid conditions the sediment yield is made up of the transport of sediment finer than 0.06 mm (washload) which is availability-limited and can be described by regression models such as the MUSLE (Section 3.4), and of coarse sand fractions which are transported at the sediment transport capacity that depends upon the local hydraulic conditions. The latter can be calculated as total load, or as bed load and suspended load which can be summed. This is best illustrated by a graph of Shen (1971) in Figure 3-2.

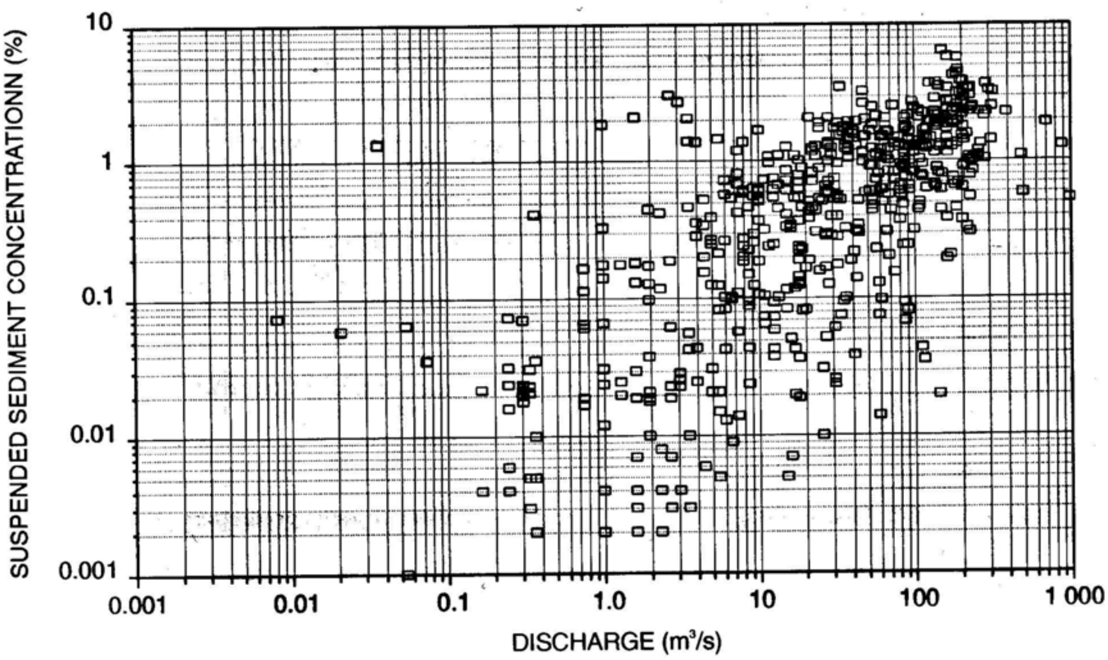

Fig 3-1: Suspended sediment data on the Caledon River, South Africa (Rooseboom, 1992)

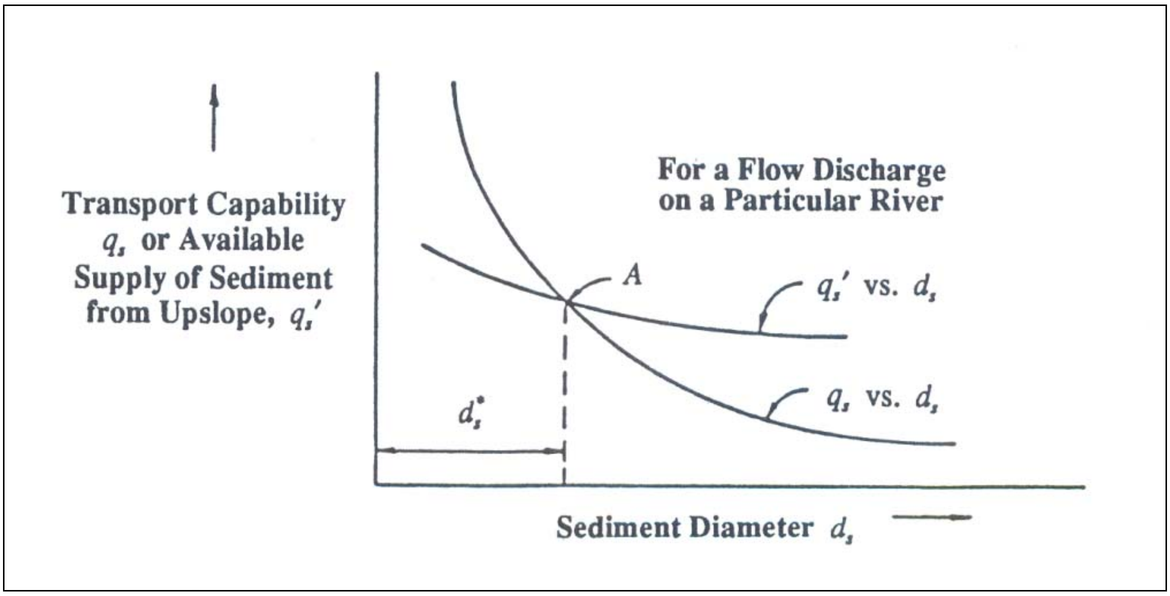

Fig 3-2: Concept of controlling sediment transport rates (Shen, 1971)

Cumulative plots of observed sediment load are very useful to determine the sediment yield which is given by the slope of the curve (Figure 3-3). Long-term changes in sediment yield can also be identified by slope changes in the graph. Figure 3-4 shows the decrease in sediment yield of the Lower Orange River, South Africa, due to land degradation by sheep farming and a decrease in availability of fine sediment. No rainfall or flow changes were observed during this period.

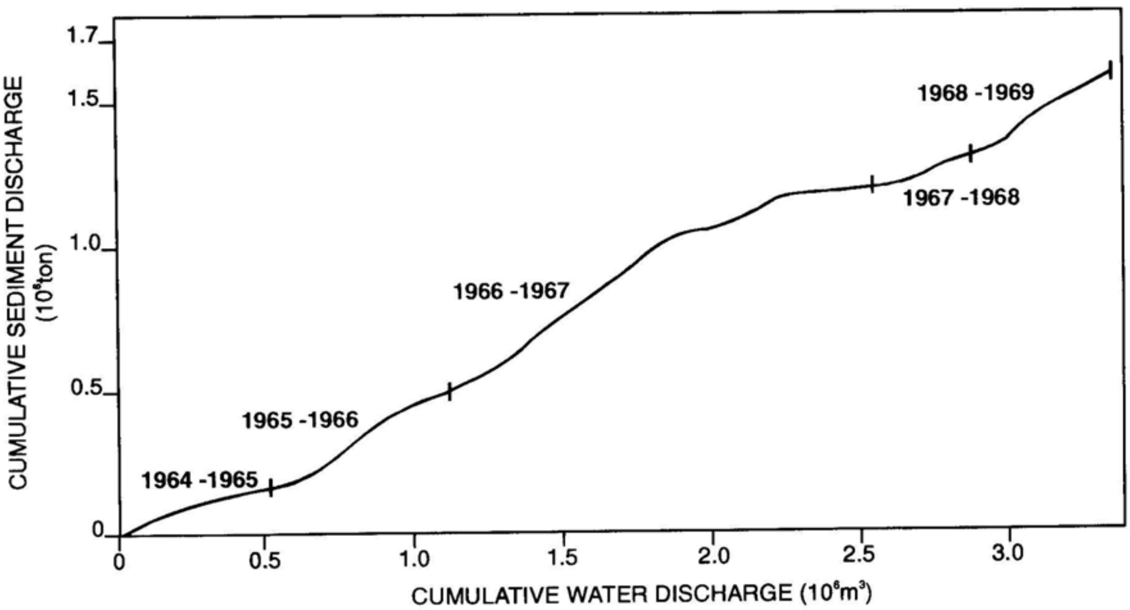

Fig 3-3: Cumulative sediment load versus discharge relationship on the Orange River, South Africa (Rooseboom, 1992)

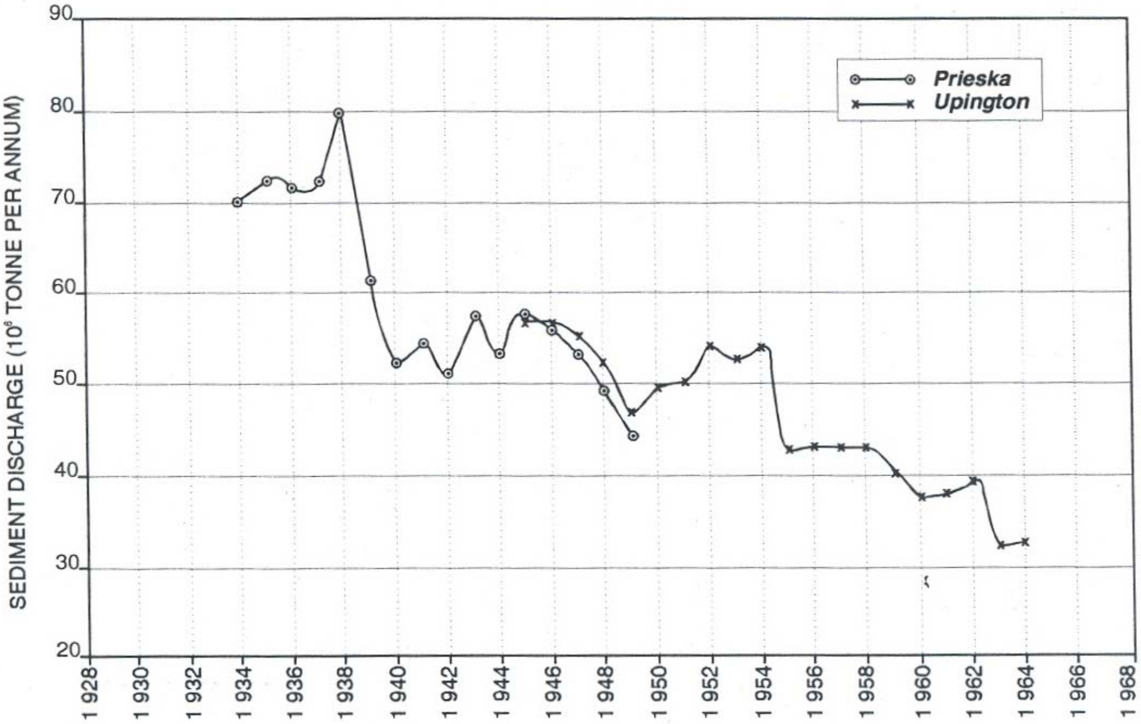

Fig 3-4: Observed sediment loads plotted for Lower Orange River, South Africa (Rooseboom, 1992)

The most reliable method of obtaining data to determine the sediment yield is river water sampling, which should include both bed load and suspended sediment sampling. The frequency of sampling should be at least daily, but the sampling should be more frequent during floods. Discharge measurement should also be carried out to determine the total load (ton/s). Sediment sampling is problematic in the remote areas in Africa and often dangerous because of crocodiles and hippos. In the semi-arid regions field research has indicated that due to the fine sediment in suspension, the vertical and lateral suspended sediment distribution is quite uniform and therefore a grab sample by hand from the river bank taken 0.3 m below the surface was found to represent the river sediment (of fine particles) concentration quite well. Bed sediment loads are difficult to obtain, especially owing to too high flow velocities and large bed dunes. Field tests indicated that a factor of 1.25 applied to a single suspended sediment grab sample takes into account the bed load and non-uniformity in suspended load across the river, and provides a realistic estimate of the total load at relatively low cost. The bed load component can also be calculated, but a stable section is required and bed roughness is often difficult to determine without detailed hydraulic investigation.

The correlation between observed discharge and sediment loads in semi-arid areas is usually poor, but can be improved by taking into account the beginning and end of the wet season, rising and falling stages of the hydrograph and the hour of the day of sampling, for example in thunderstorm regions where storms usually occur in the late afternoon hours. In future research the role of wind erosion should also be taken into account in more detail.

In semi-arid conditions the minimum record length required is 5 years, when data were found to converge to the long-term mean sediment yield (Figure 3-5). Wet and dry climatic cycles each often have a 7 to 11 year duration and should be accounted for in the sampling record period. It is important to include large infrequent floods in the record, since observations have indicated that floods with recurrence intervals of the order of 1:50 years in a single storm produce between 8 to 13 times the mean annual sediment yield.

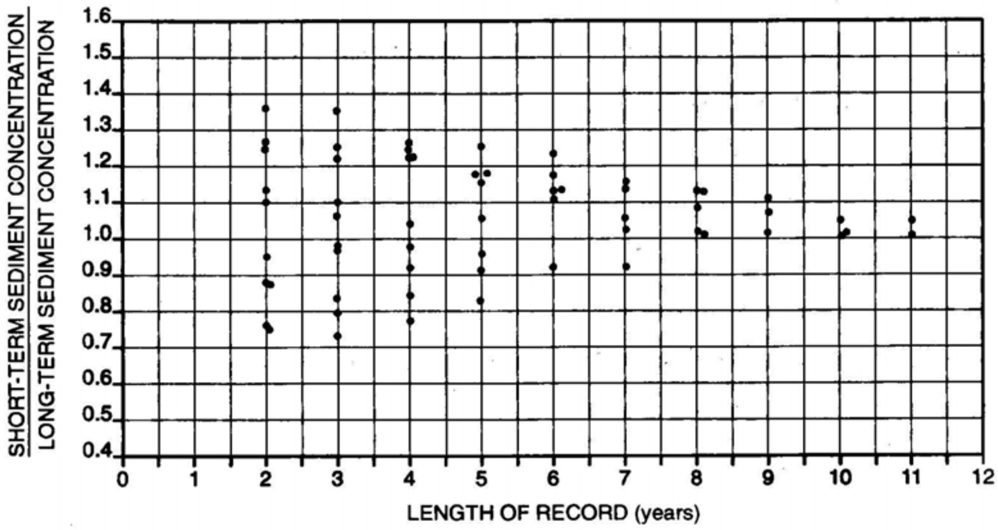

Fig 3-5: Short-term effective sediment concentration compared with long-term concentration on the Orange River (Rooseboom, 1992)

The sediment yield is usually expressed in t/km^2.year and is obtained by integrating the discharge-sediment load relationship over a long term flow record of say 40 years. The catchment area (A) is the effective area contributing to the runoff and sediment yield (e. g. downstream of large dam).

Turbidity meters or OBS meters can be employed in rivers, but usually have the disadvantage that the maximum range is only in the order of 4000 NTU. They need field calibration by suspended sediment sampling, require maintenance to prevent clogging by weeds and need power supply. Their benefit is continuous logging of point sediment concentrations, which is very important for the understanding of river sediment dynamics and for obtaining good correlation between turbidity and sediment concentration. These meters should have an upper range limit of at least 30000 mg/ℓ to 50000 mg/ℓ, in order to sample medium to large floods in semi-arid conditions.

3.2 Reservoir basin surveys

In semi-arid regions the storage capacity of a reservoir is usually in the order of the mean annual runoff, and the reservoirs therefore trap about 97% of the sediment yield. The volume of sediment deposited in the reservoir can be determined by reservoir basin surveys, say every 10 to 15 years. This volume has to be converted to a 50 year volume (V_{50}) by using an empirical equation proposed by Rooseboom (1992):

$$\frac{V_t}{V_{50}} = 0.376 \ln \frac{t}{3.5} \qquad (3.2\text{-}1)$$

With V_t = Volume of sediment after t years.

After 50 years the sediment density in reservoirs has generally been found to be about 1.35 t/m^3. The sediment yield can be determined from equation 3.2-2.

$$\text{Sediment yield (t/km}^2\text{.a)} = \frac{V_{50} x 1.35}{A x 50} \qquad (3.2\text{-}2)$$

With A = catchment area (km^2)

The following recommendations should be considered:

- Records longer than 20 years are preferable owing to the consolidation of the sediment deposits.
- Basin surveys should not be too frequent considering the limited vertical accuracy of the survey (at best 50 mm).
- Ultrasonic or sonar survey equipment to determine the bed level should be calibrated in the field for different sediments and stages of consolidation.
- Above water surveys by helicopter/fixed wing aircraft can be executed by laser but should include the backwater area during floods above full supply level (FSL), since as much a 30% of the sediment could be above FSL. In South Africa surveys are carried out to 2m above high flood level (HFL).
- The period between surveys should be representative taking into account the occurrence of flood peaks and the recurrence intervals of the floods.
- Survey of fixed cross sections is preferred so that the triangulation is carried out in the same way for all surveys when the DTM is created.

3.3 Regional sediment yield maps and statistical approach

It is often necessary to determine the sediment yield in large ungauged catchments and a regional sediment yield map or approach could be quite useful. Initially such maps have been plotted by extrapolation from observed catchment sediment yield data, in some cases also considering characteristics such as catchment area and rainfall. A statistical regional approach was also proposed by Rooseboom et al. (1992), in which a standardized yield was determined for each region, and could be multiplied by a factor that accounts for the probability of exceedance (median = 1) for the region, as well as low, medium and high sediment catchment areas based on a soil erosivity index considering soil, rainfall, slope, land use, etc.

Sediment yield prediction for large catchments based on a regionally calibrated sediment transport equation (total input stream power) considering the bed slope, sediment characteristics representing low, medium and high erosion catchment areas correlated to sediment size, the catchment area, and an effective discharge taken as the 1:10 year flood in semi-arid conditions was also found to have a fundamental basis and gave relatively good predictions. Since the highest sediment loads are transported during floods, with data indicating a single major flood can transport 8 to 13 times the mean annual sediment load, it is clear that a variable such as the 1:10 year flood would give a good correlation with long-term sediment yield. Further research is however required to validate this approach.

The above methods describing direct measurements or empirical regional approaches using GIS are far from process-based modelling, but probably give the best answers for predicting long-term sediment yields in large catchments (> 2000 km^2) because they take into account the variability of field data and avoid the many unknowns of physically-based models.

3.4 Regression type models

3.4.1 Modified universal soil loss equation (MUSLE)

The Universal Soil Loss Equation is the most widely used regression model for predicting soil erosion. It is an empirical formula for predicting soil loss caused by both sheet and rill erosion. The equation was developed from over 10000 plot-years of runoff and soil-loss data, collected since 1930 on experimental plots of agricultural land in 23 states of the United States.

Williams and Berndt (1972) recognized that application of the USLE is limited to soil loss and developed another procedure for computing sediment yields from catchments. The method determines sediment yield based on single storm events. However, MUSLE can also be used to calculate the sediment yield from the land surface on an annual basis rather than on a single storm event. This is accomplished by determining the soil loss for events of varying return periods e.g. of 2, 10, 25, 50 and 100 years, and by weighting based on the incremental probability to obtain a weighted storm average, which is then multiplied by the ratio of annual water yield to an incremental probability-weighted water yield. Long term integration of storm events and sediment transport can also be achieved by incorporating MUSLE in a hydrological model (See Section 3.4.2). MUSLE is a method which is generally applicable as a predictor of wash load and in semi-arid conditions it is more appropriate to use it than the USLE method.

With the development of hydrological models to simulate rainfall-runoff processes in larger catchments, the USLE and later MUSLE methodology were incorporated directly, initially ignoring to a large extent the completely different way in which a small catchment (plot scale) responds in terms of soil erosion and sediment yield as compared to farm scale and to large catchment scale (> 2000 km^2), and hence the sediment yields in large catchments were often overestimated.

Sediment yields are modelled on a day-by-day basis by activating the Modified Universal Soil Loss Equation, MUSLE (Williams, 1975). This version of the equation overcomes the inability of the standard USLE equation to directly determine soil loss estimates for individual storm events, and eventually eliminates the need to determine sediment delivery ratios which were used by the USLE to estimate the proportion of eroded soil which leaves the catchment.

The MUSLE sediment yield module uses factors that characterise physical conditions on the surface of a catchment as input information. Event-based sediment yield is calculated from

$$Y_{sd} = \alpha_{sy} (Q_v.q_p)^{\beta_{sy}} \; K.LS.C.P.$$

where	Y_{sd}	=	sediment yield from an individual stormflow producing event (ton)
	Q_v	=	stormflow volume for the event (m^3) from the area under study, i.e. the catchment, subcatchment or land use class
	q_p	=	peak discharge (m^3.s^{-1}) for the event
	K	=	soil erodibility factor
		=	rate of soil loss per rainfall erosion index unit
		=	f(soil texture, organic matter, structure, permeability, antecedent soil moisture condition)
	LS	=	slope length and gradient factor
		=	f(gradient)
	C	=	cover and management factor
		=	f(vegetation height, canopy cover, litter/mulch, surface roughness)
	P	=	support practice factor
		=	f(slope, conservation practices)
	α_{sy}, β_{sy}	=	location specific coefficients.

According to Simons and Sentürk (1992), the MUSLE coefficients α_{sy} and β_{sy} are site-specific, and hence must be determined for specific catchments in specific climatic regions. Kienzle and Lorentz (1993) report that very little research has been undertaken on calibrating these coefficients. Default values of 8.934 and 0.56 for α_{sy} and β_{sy} respectively, have been used in Southern Africa. Having been originally calibrated for selected catchments in the US by Williams (1975), these values for α_{sy} and β_{sy} were adopted extensively with varying degrees of success (Williams and Berndt, 1977; Williams, 1991; Kienzle *et al.*, 1997).

Conservation practices have a reduction effect on overall soil loss. Factors representing the effects of support practices can be estimated from Table 3.1 in conjunction with slope and farming practices.

Table 3.1 Conservation practice values for contour tilled lands and lands with contour banks (Wischmeier and Smith, 1978)

Land Use	Land Slope (%)	Support Practice Factor
Cultivated lands (subsistence and large-scale irrigated agriculture)	1 - 2	0.4
	3 - 8	0.5
	9 - 12	0.6
	13 - 16	0.7
	17 - 20	0.8
	21 – 25	0.9
Pastures and communal rangelands	All	1.0

3.4.2 Hydrological modelling incorporating MUSLE in ACRU

The ACRU model (Agrohydrological modelling system of the Agricultural Research Unit, South Africa) applies the MUSLE routine in subcatchments and a hydrological rainfall-runoff model routes the flow and sediment through the catchment.

The *ACRU* model was used to simulate daily sediment loads for the Mbuluzi catchment, Swaziland, for each of the 40 subcatchments (Figure 3-6) for the period 1945-1995 (Dlamini and Schulze, 2002).

Mean annual sediment yield values are presented in Figure 3-6. For the 40 subcatchments, they ranged from 59 to 9600 t/km^2.a. The highest (> 2000 t/km^2.a) values of sediment yields were simulated in the northeastern part of the catchment where the subcatchment has the highest average slope at 16%, and is occupied by rural communities with more than 20% of the land under subsistence agriculture, the remainder being grazed and browsed bushlands and forests. Other high mean annual sediment yields were simulated in the upper-middle parts, with 1709 t/km^2.a. This region is also predominantly rural, with subsistence agriculture being the main farming activity, while all the unimproved grasslands (which cover more than 70% of the land) are used as communal pastures. During fieldwork, land with relatively steep slopes was found to be cultivated. Bare patches of land, badlands (gullies) and livestock and human pathways, which are sources of sediments, were also observed in the rangelands.

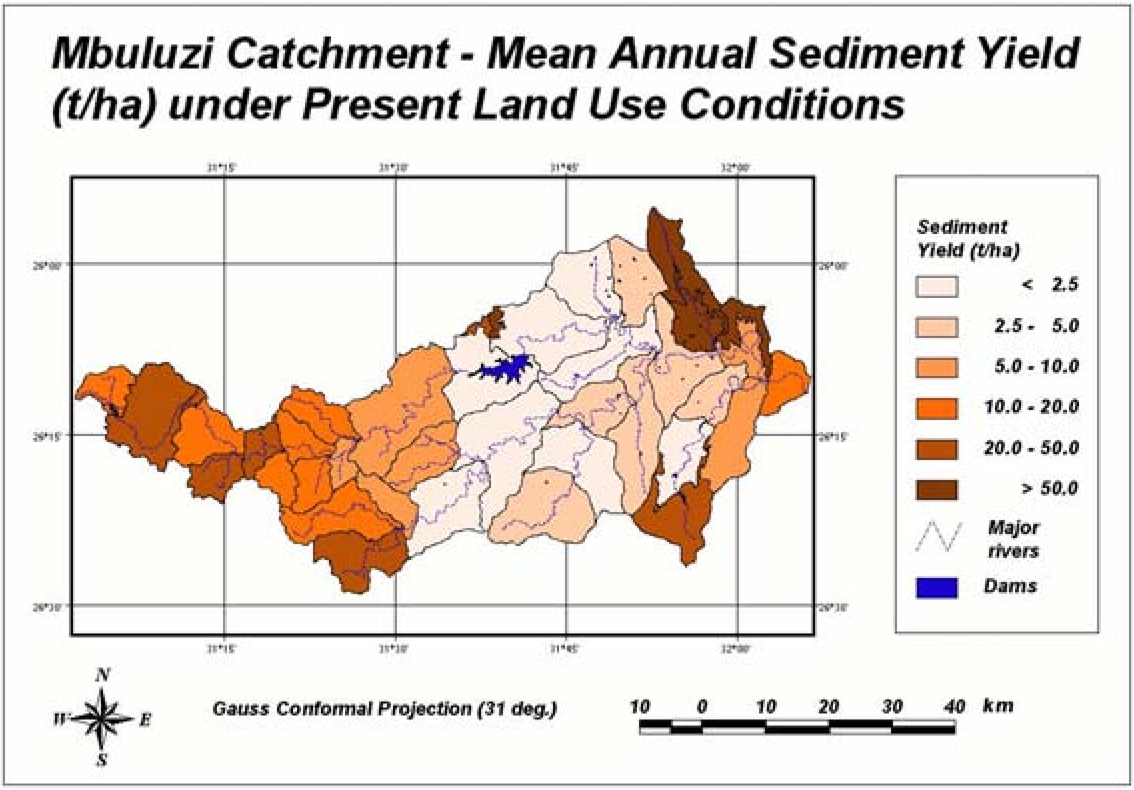

Fig 3-6: Simulated mean annual sediment yields ($t.ha^{-1}$) in the Mbuluzi catchment (Dlamini and Schulze, 2002)

In this example the simulated sediment yield was not validated against field data (none available), but the predicted values seem to be relatively high. The strength of such a model is however to analyse the relative importance of different land uses and of possible rehabilitation measures. Two questions can be addressed by model simulations:

a) How do sediment yields from different land uses compare with one another?
b) What is the impact of veld degradation or rehabilitation on sediment yields?

A comparison of sediment yields simulated under different land uses, using one of the subcatchments by way of example (Figure 3-7), indicates that subsistence agriculture and rangelands, i.e. grasslands in poor hydrological condition, produce the highest and second highest sediment yields respectively, while land under forest and rehabilitated grasslands generate the least sediment yields. The sediment yields under subsistence agriculture are highest in November, which is the ploughing and planting month for maize (the crop most commonly grown by rural Swazis), when the soil is exposed. Of note is that sediment yields simulated in the grassland in poor hydrological condition are higher than those of subsistence agriculture between February and March. This is a consequence of the mature stage maize has then reached, plus the improvement in ground cover following the growth of weeds, coinciding with the continued grazing and degradation of the grasslands (Dlamini and Schulze, 2002).

It is common practice in the rural areas to allow livestock to freely roam the maize fields after harvesting between April and the beginning of the planting period, leaving rangelands to recover. Hence the higher sediment yields under the subsistence agriculture over that period.

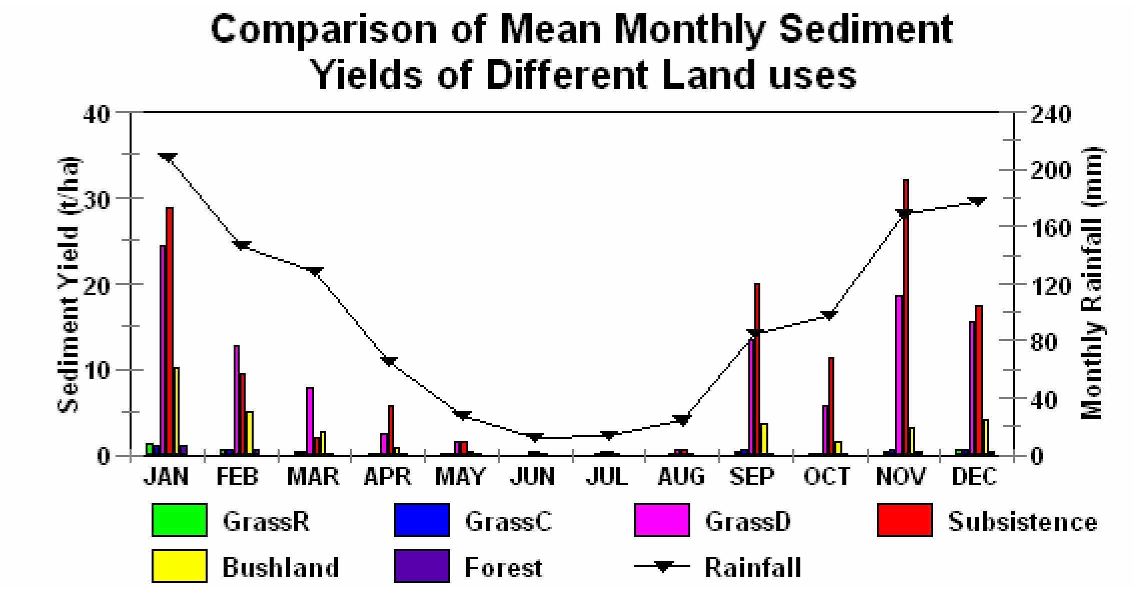

Fig 3-7: Comparison of sediment yields simulated under different land uses in Subcatchment 6. Under grasslands, R designates rehabilitated (i.e. well managed) conditions, C current and D degraded (overgrazed) conditions (Dlamini and Schulze, 2002)

One objective of this study (Dlamini and Schulze, 2002) was to assess the effects of land use management on sediment yields. The mean annual sediment yields were reduced in all the subcatchments after replacing those areas of the present land cover which can be grazed with a grass cover in good hydrological condition. High reductions ranging from 500 to more than 2500 t/km^2.annum were found in subcatchments in the upper-middle and upper sectors of the catchment. Note that these are the subcatchments that presently produce high sediment yields (Figure 3-8).

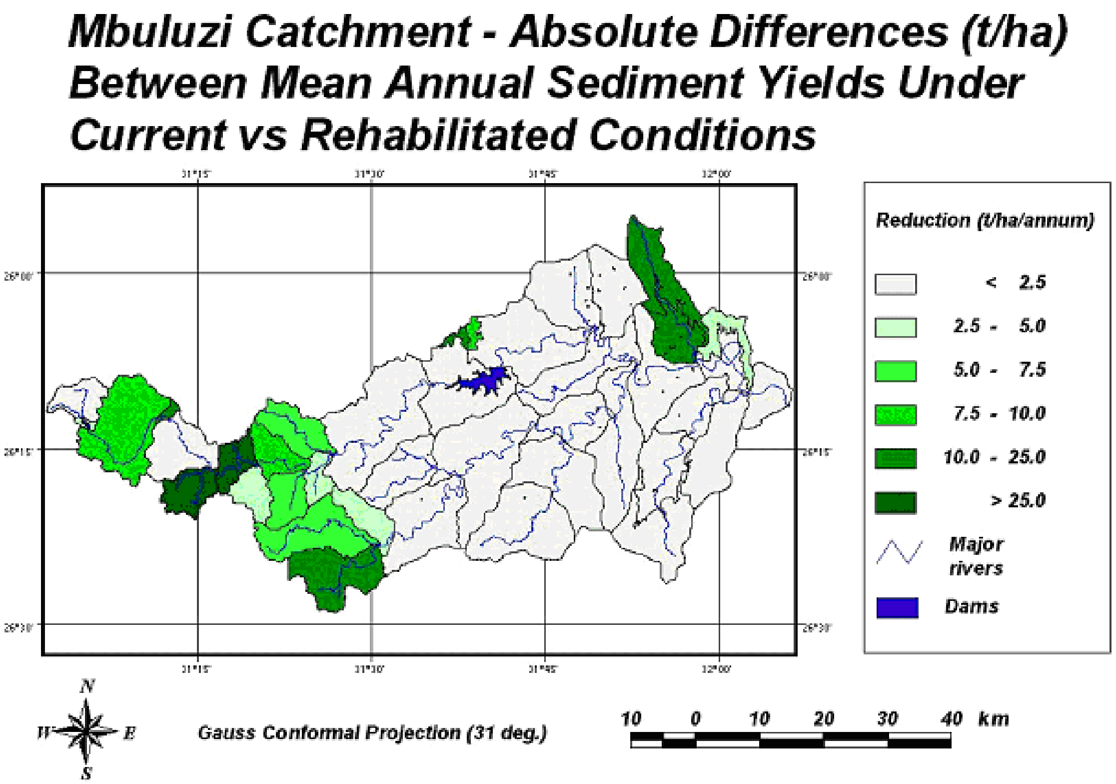

Fig 3-8: Absolute (t/ha) differences between simulated mean annual sediment yields under rehabilitated vs degraded conditions (Dlamini and Schulze, 2002)

3.5 Physically-based erosion and sediment yield models

3.5.1 General

Physically-based, spatially-distributed modelling systems have particular advantages for the study of basin change impacts and applications to basins with limited records. Their parameters have a physical meaning (e.g., soil permeability and sediment size distribution) and can be measured in the field.

Disadvantages of physically-based models include heavy computer requirements, the need to evaluate many parameters (with associated problems of representation at different spatial scales and uncertainty) and a complexity which implies a lengthy training period for new users.

Two state-of-the-art models are described here: the WEPP model (Flanegan and Nearing, 1995) (USA) and SHETRAN (UK) (Norouzi et al., 2004). These models differ from the ACRU model with the MUSLE routine in that they use a much more detailed mathematical description of the sediment dynamics and routing processes in the catchment.

3.5.2 The WEPP Model

WEPP is a continuous simulation model, and can use either observed or generated climatic inputs to drive the runoff and erosion processes. Critical components of WEPP are the infiltration and runoff computations. Depressional storage is estimated as a function of random roughness and slope steepness. When the rainfall rate exceeds the infiltration rate, rainfall excess is be computed. Runoff is the total rainfall excess minus any reduction due to the surface depressional storage.

The peak runoff rate is a very important parameter in WEPP, as it is used in calculations to estimate flow depth and ultimately flow shear stress. WEPP uses either a semi-analytical solution of the kinetic wave model or an approximation of the kinetic wave model to determine the peak runoff rate. Runoff rate, rill roughness and rill channel characteristics are used with the Darcy-Weissbach equation to estimate flow depth and hydraulic radius. Sediment transport capacity is computed using a simplified function of shear stress raised to the 3/2 power, times a coefficient that is determined through application of the Yalin equation at the end of the slope profile.

The WEPP model uses a steady-state sediment continuity equation to predict sediment load down a hillslope profile:

$$\frac{dG}{dx} = D_f + D_i \qquad (3.5\text{-}1)$$

where G is sediment load ($kg.s^{-1}.m^{-1}$), x is distance downslope (m), D_f is rill erosion rate ($kg.s^{-1}.m^{-2}$), and D_i is interrill sediment delivery rate ($kg.s^{-1}.m^{-2}$) . Interrill sediment delivery to rills is predicted in WEPP using the following equation:

$$D_i = K_{iadj} I_e O_{ir} SDR_{RR} F_{nozzle} \frac{R_s}{w} \qquad (3.5\text{-}2)$$

where K_{iadj} is the adjusted interrill erodibility factor ($kg.s.m^{-4}$), I_e is effective rainfall intensity, ($m{\cdot}s^{-1}$) O_{ir} is the interrill runoff rate, ($m{\cdot}s^{-1}$), SDR_{RR} is a sediment delivery ratio that is a function of random roughness, row side-slope and the interrill particle size distribution, F_{nozzle} is an adjustment factor to account for sprinkler irrigation nozzle impact energy variation, R_s is the rill spacing (m), and w is the rill width (m). Rill erosion rate may be either positive in the case of detachment or negative in the case of deposition. Rill detachment in WEPP is predicted when the flow sediment load is below transport capacity, and flow shear stress acting on the soil exceeds critical shear stress. In that case, D_f is predicted with:

$$D_f = K_{radj}\,(\tau - \tau_{cadj})(1 - \frac{G}{T_c}) \qquad (3.5\text{-}3)$$

where K_{radj} is the adjusted rill erodibility factor ($s \cdot m^{-1}$), τ is flow shear stress (Pa), τ_{cadj} is adjusted critical shear stress of the soil (Pa), G is sediment load in the flow ($kg.s^{-1}.m^{-1}$), and T_c is flow sediment transport capacity ($kg.s^{-1}.m^{-1}$).

Other model components include a soil component to adjust roughness, infiltration, and erodibility parameters as affected by tillage and consolidation, a plant growth component to provide daily values of crop canopy, biomass, and plant water use, and a daily water balance to determine the impacts of soil evaporation, plant transpiration, infiltration, and percolation on soil water status. Crop residue levels are also updated daily, with adjustments for decomposition as well as the impacts of tillage or other management operations. WEPP contains components to estimate frost, thaw and snow depths, as well as snow melt runoff in regions that experience freezing temperatures. Additionally, the model can be used to determine the impact of furrow and sprinkler irrigation on soil erosion.

In watershed applications, WEPP allows simulations of groups of hillslopes, channels, and impoundments. Daily water balance, plant growth, and soil and residue status for channels are predicted identically to that on hillslopes. Channel peak runoff rates are predicted using either a modified Rational Equation or the CREAMS (Knisel, 1980) peak runoff equation. Channel erosion is estimated using a steady-state sediment continuity equation:

$$\frac{dq_{sed}}{dx} = D_L + D_F \qquad (3.5\text{-}4)$$

where q_{sed} is the sediment load in the channel ($kg \cdot s^{-1} \cdot m^{-1}$), x is distance down the channel (m), D_L is lateral inflow of sediment along the channel ($kg \cdot s^{-1} m^{-2}$), and D_F is detachment or deposition by flow in the channel ($kg \cdot s^{-1} m^{-2}$). For a channel in an active detachment mode that has not reached a nonerodible layer, a rectangular channel is assumed and the erosion rate is:

$$E_{ch} = w_c K_{ch}\,(\tau_{ave} - \tau_{cr}) \qquad (3.5\text{-}5)$$

where E_{ch} is the soil loss per unit channel length ($kg.s^{-1}.m^{-1}$), w_c is channel width (m), K_{ch} is a channel erodibility factor ($s.m^{-1}$), τ_{ave} is average channel flow shear stress acting on the soil (Pa), and τ_{cr} is critical shear stress of the channel soil (Pa).

If sediment load of all particle types is larger than flow sediment transport capacity, then sediment deposition in the channel is predicted using:

$$D_F = \frac{v_f}{q_w}\,(T_c - q_{sed}) \qquad (3.5\text{-}6)$$

where v_f is particle fall velocity ($m.s^{-1}$), q is flow discharge per unit width ($m^2.s^{-1}$) and T_c is channel flow sediment transport capacity ($kg.s^{-1}.m^{-1}$). For cases in which sediment load is near transport capacity, shifting of transport capacity from particle classes with excess to those with a deficit is predicted.

WEPP ignores soil saturation at the foot of a hillslope due to saturation-excess overland flow, and may well fail to predict important erosional features in a catchment. The model also cannot simulate gully erosion which could be an important component in semi-arid regions.

3.5.3 SHETRAN

a) Model description

SHETRAN (Norouzi et al., 2004) is a physically-based, spatially-distributed, integrated surface/subsurface modelling system for water flow, sediment transport and contaminant migration in river basins, which has been developed at the Water Resource Systems Research Laboratory (WRSRL), Department of Civil Engineering, University of Newcastle upon Tyne. Its original basis was the Système Hydrologique Européen (SHE) hydrological modelling system.

Compared with more traditional modelling approaches SHETRAN has particular advantages in representing distributed responses at catchment scales from less than 1 km^2 to 2000 km^2, in predicting the impacts of land use and climate change, in incorporating landslide and gully erosion and in exploring issues such as scale effects and validation techniques which are at the forefront of physically-based modelling research.

SHETRAN is a general, physically-based, spatially-distributed modelling system: that is, it can be used to construct and run models of all or any part of the land phase of the hydrological cycle (including sediment and contaminant transport) for any geographical area. It is physically-based in the sense that the various flow and transport processes are modelled either by finite difference representations of the partial differential equations of mass, momentum and energy conservation, or by empirical equations derived from experimental research. The model parameters have a physical meaning and can be evaluated by measurement. Spatial distributions of basin properties, inputs and responses are represented on a three-dimensional, finite-difference mesh. The channel system is represented along the boundaries of the mesh grid squares as viewed in plan.

The typical processes modelled by the SHETRAN hydrological component are:

- Interception of rainfall on vegetation canopy (Rutter storage model)
- Evaporation of intercepted rainfall, ground surface water and channel water; transpiration of water drawn from the root zone (Penman-Monteith equation or the ratio of actual to potential evapotranspiration as a function of soil moisture tension)
- Snowpack development and snowmelt (temperature-based of energy budget methods)
- One-dimensional flow in the unsaturated zone (Richards equation)
- Two-dimensional flow in the saturated zone (Boussinesq equation)
- Two-dimensional overland flow; one-dimensional channel flow (Saint Venant equations)
- Saturated zone/channel interaction, including an allowance for an unsaturated zone below the channel
- Saturated zone/surface water interaction

The basic erosion and sediment yield component consists of subcomponents accounting for soil erosion by raindrop impact, leaf drip impact and overland flow, channel bed and bank erosion by channel flow, and sediment transport by overland and channel flow.

The sediment processes modelled and the equations used to describe them are:

- Soil erosion by raindrop impact, leaf drip impact and overland flow.
- Two-dimensional total load convection in overland flow by size fraction, including input to the channels; deposition and resuspension of sediments in overland flow (mass conservation equation incorporating Engelund-Hansen total load and Yalin load transport capacity equations).

- One-dimensional convection of cohesive and noncohesive sediments in channel flow by size fraction; deposition and resuspension of noncohesive sediments in channel flow; channel bed erosion by channel flow (mass conservation equation incorporating Ackers-White and Engelund-Hansen transport capacity equations)
- Landslide erosion and sediment yield component
- Gully erosion

The SHETRAN subcomponent is applied at each grid square to generate a response (e.g., overland flow, phreatic surface rise, soil erosion). These responses interact and both surface and subsurface waters, and surface sediments, are routed from square to square. Eventually these products reach the river system and are routed towards the basin outlet. Model outputs may be obtained for any part of this procedure on a spatially and temporally distributed basis. They may include time-varying records of phreatic surface level, snowpack depth, overland flow depth, soil erosion or any other variable at any grid square or channel link within the basin.

b) Parameter uncertainty and model validation

The model parameters or functions to which the simulation results are typically most sensitive in full basin simulations are: saturated zone hydraulic conductivity; unsaturated zone hydraulic conductivity; Strickler resistance coefficient for overland flow; soil retention curve (the relationship between tension and moisture content); the relationship between the ratio of actual to potential evapotranspiration and soil moisture tension; and, the soil erodibility coefficients.

c) Parameter scale effects

Within each model grid square, each physical characteristic is represented by one parameter value. As long as the grid square is small compared with the distances over which there is significant spatial variability in catchment properties and hydrological response, this does not compromise the model's ability to represent local variations in response. However, as grid scales increase, the local spatial variability in properties and response becomes subgrid (one parameter value describing physical characteristics per grid square not sufficient). There are then difficulties in applying the equations of small scale physics which make up SHETRAN and evaluating their parameters at the grid scale.

Studies with SHE, SHETRAN and ANSWERS (another distributed model) suggest that the same model parameter values can be applied at both plot (1-100 m^2) and microbasin (order 1 ha) scales, using small model grid spacing (20 m or less) and having a good availability of field data at one's disposal. For larger basins, scale effects in evaluating saturated zone conductivity appear not to be significant, or at least to be masked by uncertainty in parameter evaluation, as long as basin topography is subdued and there is a general homogeneity of land use, soil characteristics and hydrological response within the basin. Applications of the SHE to large basins in India (area 800-5000 km^2) and to the Cobres basin in Portugal (area 701 km^2) suggest that conductivities evaluated at the point or small scale can be successfully applied with a model grid spacing of 2 km. Figueredo (1998) similarly found no evidence of a scale effect when modelling a 137 km^2 basin in northeast Brazil, although in this case the basin did not typically have a saturated zone in the soil column. For the dissected terrain of the Draix basins, Bathurst et al. (1998) concluded that any scale effects which may distinguish simulations at the scales of 0.133 and 86 ha were small enough to be masked by uncertainty in parameter evaluation. However, for basins of 200-2000 km^2 with hilly terrain, unpublished simulation results suggest that the saturated zone conductivity may indeed increase as the grid spacing rises to 1 or 2 km.

For overland flow resistance, the picture is less clear. If there is any scale dependency the effect does not appear to be large and other factors such as the type of ground roughness (perhaps determined by land use) may be more important. No scale dependency has yet been observed in the soil erodibility coefficients.

3.6 Disadvantages of physical process simulation models

Physical process simulation models have several disadvantages compared to the regression-type models such as the MUSLE. These disadvantages are related mostly to the increased complexity of the physical process models. Firstly, the models of large catchments are computationally extremely "heavy".

Another disadvantage is that data requirements are more extensive because of increased complexities. However, in some ways data requirements are simplified for physical process models in that the necessary data are more easily measured and identified because of the physical process basis. Data requirements for regression models are often much more subjective and the parameters often harder to relate to observable and measurable quantities.

The complexity of physical process models requires broad knowledge of erosion and sedimentation, and watershed knowledge in general. Without adequate background, it is doubtful that physical process models can be applied to achieve accurate results. Most regression models require less knowledge since the procedures for their applications are usually rigid. However, application of a regression model without proper training and understanding can produce invalid results.

3.7 Evaluation studies of soil erosion models

Validating the ability to model spatial and temporal variations in erosion and sediment fluxes requires spatially and temporally varied measurements. Over the last decade or so, developments in field instrumentation and advances in data collection techniques have provided the capability for satisfying much of this demand.

A series of studies have been conducted to compare erosion model predictions of soil loss to measured data. These include studies on WEPP the USLE and RUSLE. Risse et al. (1993) applied the Universal Soil Loss Equation (USLE) to 1700 plot years of data from 208 natural runoff plots. Average observed soil loss on an annual basis was 3.51 kg/m^2. Using the USLE, annual values of predicted soil loss averaged 3.22 kg/m^2 with an average magnitude (absolute value) of error of 2.13 kg/m^2, or approximately 60% of the mean. Rapp (1994) applied the RUSLE model to the same set of data as Risse et al., and annual values of predicted soil loss averaged 3.16 kg/m^2. The average magnitude (absolute value) of error was not reported, but it is apparent that the two models performed similarly overall in terms of soil loss prediction. Zhang et al. (1996) applied the WEPP computer simulation model to 290 annual values and obtained an average of 2.18 kg/m^2 for the measured soil loss, with an average magnitude of error of 1.34 kg/m^2, or approximately 61% of the mean. In both cases the relative errors tended to be greater for the lower soil loss values. All three studies were conducted without model calibration. Model input parameters were not adjusted from initial default values for the specific data used in the comparisons.

What is reported above is obviously a very "broad brush" picture of the performance of the three erosion models, but in essence, the results indicate that for the prediction of soil loss, the three models appear to perform approximately on the same level of accuracy. However, there are a couple of important points to be considered. In the first place, all three models do predict soil loss, but only the WEPP model is specifically designed to predict sediment yield. Thus if prediction of average soil loss on the eroding portion of a hillslope is the goal, one might conclude from the studies that any of the three models work equally well. However, if one needs to know the deposition rates in the toe-slope of the hill, how much sediment might be transported off-site, sediment load from a channelised area, or the distribution of erosion along the hillslope, only WEPP will provide that information.

3.8 Model Comparison

A comparison of 6 physically-based erosion and sediment yield models is presented in Table 3.2.

Table 3.2 Comparison of six physically-based erosion and sediment yield models

Model Feature	SHETRAN	ANSWERS	WEPP	EUROSEM	LISEM	ACRU
Simulation type:						
Continuous	Y	N	Y	N	N	Y
Single event	Y	Y	Y	Y	Y	Y
Basin size	<2500 km^2	<50 km^2	<2.6 km^2	Small basin	Small basin	<10000 km^2
Spatial distribution	Grid	Grid or GIS raster	grid	Uniform slope planes	GIS raster	GIS raster
Overland flow:						
Rainfall excess	Y	Y	Y	Y	Y	Y
Upward saturation	Y	N	N	N	Y	Y
Erosion process:						
Raindrop impact/ Overland flow	Y	Y	Y	Y	Y	Y
Rilling	N	N	Y	Y	Y	Y
Crusting	N	Y	N	Y	Y	Y
Channel banks	Y	N	N	Y	N	N
Gullying	Y	N	N	N	N	N
Landsliding	Y	N	N	N	N	N
Output:						
Time-varying sedigraph	Y	Y	N	Y	Y	Y (daily)
Time-integrated yield	Y	Y	Y	Y	Y	Y
Erosion map	Y	Y	Y	Y	N	Y
Land use	Most vegetation covers	Mainly agricultural	Wide range of land use	Mainly agricultural	Mainly agricultural	Mainly agricultural

Y = yes; N = no

4. RIVER SEDIMENT TRANSPORT

Once sediment yields from sub-catchments have been determined, the sediment has to be routed downstream in the river system by mathematical modelling.

4.1 Hydrodynamic mathematical modelling

Accurate prediction of sediment transport should be based on reliable hydrodynamic modelling. Such a state of the art model could have the following characteristics (Basson and Rooseboom, 1997):

4.1.1 One-dimensional (1D) or 2D or 3D Models

One-dimensional models are mostly used in river applications around the world. Two-dimensional (2D horizontal or vertical) and 3D models have been developed, although they are computationally "heavy". Two-dimensional models are especially taking over from 1D models to describe the fluvial morphology. The main constraints in using a 2D model with sediment transport are often a lack of data for calibration of the model, and the uncertainty of the sedimentation processes.

Two-dimensional models can be of specific benefit when considering:

- deposition outside the main channel across the wide-open floodplains when encountered
- modelling of sediment transport through a reservoir or estuary
- Braided or meandering channel fluvial morphology (2DH)
- Density current formation (2DV or 3D)

Many 2DH models are pseudo-3D and have empirical coefficients to account for spiral flow at river bends, vertical sediment concentration distribution, etc.

4.1.2 St. Venant equations for hydrodynamic simulation

A fully hydrodynamic approach is required to describe the rapidly changing flow and bed level conditions during floods when most of the sediment is typically transported. The model must also be able to simulate the supercritical flow conditions which can be encountered. The use of daily flows in a quasi-steady modelling approach is often not suitable except in large rivers. This is because flood peaks are averaged which leads to unreliable sediment transport prediction.

The hydrodynamics described by the 2-dimensional Saint Venant equations are:

$$\frac{\partial p}{\partial t}+\frac{\partial (p^2/h)}{\partial x}+\frac{\partial (pq/h)}{\partial y}+\frac{gp\sqrt{p^2+q^2}}{C^2h^2}+gh\frac{\partial s}{\partial x}=h\frac{\partial}{\partial x}\left(E\frac{\partial (p/h)}{\partial x}\right)+h\frac{\partial}{\partial y}\left(E\frac{\partial (p/h)}{\partial y}\right) \quad (4\text{-}1)$$

$$\frac{\partial q}{\partial t}+\frac{\partial (pq/h)}{\partial x}+\frac{\partial (q^2/h)}{\partial y}+\frac{gq\sqrt{p^2+q^2}}{C^2h^2}+gh\frac{\partial s}{\partial y}=h\frac{\partial}{\partial x}\left(E\frac{\partial (q/h)}{\partial x}\right)+h\frac{\partial}{\partial y}\left(E\frac{\partial (q/h)}{\partial y}\right) \quad (4\text{-}2)$$

$$\frac{\partial h}{\partial t}+\frac{\partial p}{\partial x}+\frac{\partial q}{\partial y}=0 \quad (4\text{-}3)$$

Where:

p,q	Flux field (m^2/s)
h	Water depth (m)
t	Time (s)
g	Acceleration of gravity (9.81 m/s^2)
C	Chezy number ($m^{1/2}/s$)
s	Surface elevation (m)
E	Eddy viscosity (m^2/s)

The Saint Venant equations are solved on the curvilinear grid with a finite-volume scheme. State of the art methods from CFD are applied for the solution, which incorporates the use of a Cartesian base for the velocity field, non-staggered allocation with momentum interpolation (Majumdar et al, 1992). The SIMPLER method (Patankar, 1980) is applied for the continuity equation.

4.1.3 Secondary flow patterns at river bends

At river bends the modified flux field that transports the suspended sediment is derived from the depth-integrated flux field in the manner (DHI, 2003):

$$\begin{pmatrix} p' \\ q' \end{pmatrix} = \alpha_{01} \begin{pmatrix} p \\ q \end{pmatrix} + \alpha_{02} \frac{h}{R} \begin{pmatrix} -q \\ p \end{pmatrix} \tag{4-4}$$

α_{01} and α_{02} are functions of the distribution of momentum and sediment over the water column. The term h/R is the water depth divided by the streamline radius of curvature; the latter derived from the flow field. The modified flux field arises from the 3-dimensional character of the problem.

α_{01} modifies the streamwise convection, and represents the fact that the sediment concentration rises towards the bed, while the velocity rises towards the surface. The streamwise convection of the sediment is hence not as effective, as $\alpha_{01}=1$ would imply. A value of $\alpha_{01}=1$ is found for uniformly distributed sediment, i.e. very fine material. α_{01} is calculated from the logarithmic velocity profile and the distribution of sediment.

α_{02} represents the impact of secondary flow, and produces convection across the streamlines. α_{02} is calculated from the helical flow taken from standard theory (see e.g. Rozowskii, 1957), and the distribution of sediment.

The α_{01} and α_{02} parameters are calculated from local values of the flow velocity, flow resistance and settling velocity. The calculation is done on each morphological time-step in describing the change in sediment concentration and sedimentation in time.

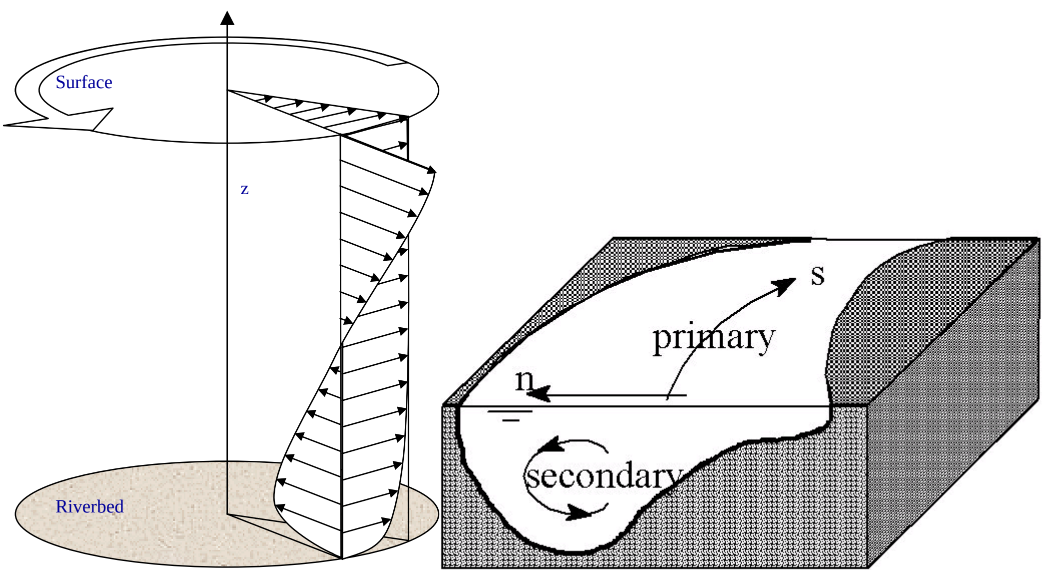

Fig 4-1: Profile functions in pseudo 3D model (DHI, 2003)

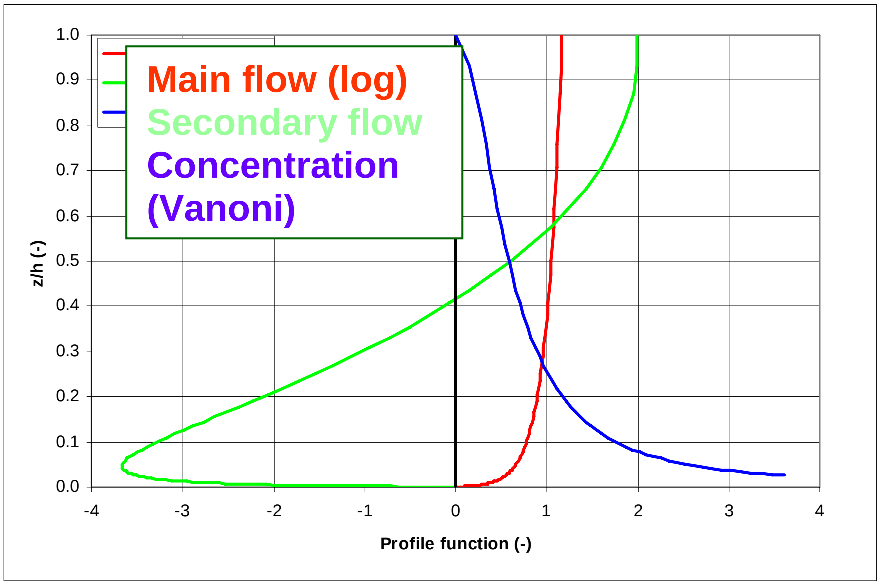

Fig 4-2: Vertical flow and concentration distribution (DHI, 2003)

An implicit scheme is applied for the Advection-Dispersion (AD) equation (also refer to section 4.2.3) in which the local availability of sediment is accounted for by limiting the erosion to not surpass the available cohesive sediment in the cell. The implicit solution of the AD equation furthermore allows for implicit updating of the local cohesive layer thickness, which is done through the source/sink terms of the equation (sediment entering the water column comes from the bed, and vice versa). The implicit AD scheme is unconditionally stable for any choice of the time-step. The dispersion in the equation originates from the profile functions, while additional dispersion can be added. Herein is only applied the dispersion associated with the profile functions (particularly important across streamlines). No additional dispersion has been added.

The models use finite differences, finite elements or finite volumes to solve the set of equations. For the grid generation an orthogonal curvilinear grid or triangular grid is recommended, which has no constraint on the cell aspect ratio.

Mathematical models should be calibrated against water level and discharge field data to account for bed roughness of alluvial beds, bed rock, riparian vegetation, tributaries, hydraulic structures such as diversion weirs, abstractions, losses, etc. Boundary conditions are typically tributary and upstream inflow time series and a water level-discharge boundary downstream. Bed roughness is the most important calibration factor and is spatially calibrated against observed flood attenuation and flood peaks, but large scale roughness during low flows is also important to consider. Measured tributary flows are often scaled up to account for unmeasured subcatchment areas and this also forms part of the calibration process.

In general the hydrodynamic models can be calibrated accurately to within 20 to 30% accuracy based on flood peaks. It should be noted that flood data are usually limited and observed flood peak discharge also has an accuracy range of say 10 to 30% typically, which is improving recently with the use of acoustic Doppler profilers using sonar for bed tracking with differential GPS for geo-referencing.

4.2 Sediment Transport Modelling

4.2.1 Sediment transport capacity of non-cohesive sediment

The basis of sediment routing in a river is the sediment transport equation used. When dealing with non-cohesive sediments where the sediment diameter > 0.03 mm, several sediment transport capacity equations are widely used. Most of these equations have a streampower basis and have been calibrated against laboratory and/or field data. These equations were derived under steady and uniform flow, uniform and equilibrium bed conditions, usually in a laboratory. Field conditions, especially during floods when most sediment is transported are usually the opposite. The accuracy of these equations vary based on who carried out the accuracy comparison and is often biased as shown in Table 4.1, but generally a 50% prediction accuracy is considered to be good, even with controlled laboratory experiments.

What is worse is that in the Table 4.1 comparison all the hydraulic parameters such as flow depth and velocity were known, while in mathematical modelling it is simulated and calibrated. Therefore, there seems to be room for improving the sediment transport prediction accuracy, but the inaccuracy cannot be blamed on the theory alone but also on the data reliability and variability. Further important aspects of sediment transport equations are that they have usually only been calibrated for sediment size ranges of say 0.07 mm to about 2 mm, while some are calibrated on gravel transport up to about 30 mm in diameter.

Mathematical modelling of alluvial beds with sand usually assumes an infinite supply of sand, and transport capacity is calculated in each time step at the boundaries feeding into the river. Sand is however not always available, such as at bed rock reaches and this fact should be taken into account.

4.2.2 Non-uniform sediment modelling

Most sediment transport equations were calibrated on near uniform sediment sizes. In reality, however, sediment particles in a river bed are far from uniform in diameter. In such cases a multi-fraction approach is followed for non-cohesive sediment fractions, the transport capacity of each fraction is calculated and possible re-entrainment from the bed. This approach assumes that different fractions in the bed react independently with no interrelationship. Yet, the coarser particulars could hide smaller particles, and fine particles with a small number of large particles could mean the larger particles are exposed. In some cases hiding could lead to armouring where large particles cover the finer particles making it impossible for the small particles to be entrained, except when the larger particles are re-entrained. Although several theories have been developed to describe sediment hiding, armouring and exposure in non-uniform beds, they are seldom used in sediment transport simulations.

In the above discussion no mention was made of even more non-uniform sediment distributions namely cohesive-non-cohesive sediment mixtures and this is because in sediment transport modelling the sediment is usually either treated as non-cohesive or cohesive, but not both. Even in semi-arid conditions where up to 80% of the sediment transported during a flood consists of clay and silt fractions, this is often ignored since most mathematical models cannot deal with clay/silt and sand fractions. In any case, it is argued that the cohesive sediments do not affect the river geomorphology and are transported through to the ocean. Only in depositional zones such as lakes, reservoirs or estuaries, it is further argued, that it may be important to include cohesive sediment fractions. The prediction of fine sediment transport should however be included and is important right through the river system: cohesive sediment deposits on the floodplain and affects bank stability. It also affects critical conditions for erosion and when the cohesive sediment fraction in the bed exceeds only 7%, the bed reacts as cohesive with a different bed roughness. A mathematical model therefore has to keep book of the bed sediment composition, which adjusts with sediment erosion and deposition.

To come back to the multi-fraction approach described earlier, it often happens that the fine sediment, the so-called washload, is transported through the river without deposition. It is however essential in reservoir modelling that sediment transport calculations are carried out per size fraction in order to model the sorting process and related non-cohesive and cohesive deposits through the reservoir.

Bank stability models can also be incorporated in 2D models based on the geotechnical stability of the bank and is important at river bends.

Table 4.1 Non-cohesive sediment transport equation accuracy (Basson and Rooseboom, 1997)

White et al., (1975) compared eight formulae using 1000 flume and 260 field measurements. The discrepancy ratio X_{calc}/X_{obs} was plotted against the dimensionless grain size ($\rho_s q_s / \rho q$) and the percentages within the 0.5 to 2 range were as follows:

Formula	**% in $0.5 \leq X_{calc}/X_{obs} \leq 2$ ranges**
Ackers and White (1973)	68
Engelund and Hansen (1967)	63
Rottner (1959)	56
Einstein (1950) (total load)	46
Bishop et al., (1965)	39
Toffaletti (1968)	37
Bagnold (1966) (total load)	22
Meyer-Peter and Müller (1948)	10

The laboratory data include particle sizes from 0.04 to 4.94 mm and field data from 0.095 to 68 mm.
The comparison of formulae by Yang and Molinas (1982) also used laboratory and river data encompassing mean grain sizes from 0.15 to 1.71 mm, channel widths 0.134 to 532 m, flow depths 0.01 to 15.2 m, temperature 0º to 34.3ºC, average velocity 0.23 to 1.97 m/s and slopes from 4.3×10^{-5} to 2.79×10^{-2}. The range of data is the same as given by Yang (1973) for the data from which the formula was derived. The discrepancy ratio, defined as the ratio between computed and measured values, is given as follows:

	Data		
Formula	**Lab.**	**River**	**All data**
Colby (1964)	0.31	0.61	0.34
Yang (1973)	1.01	1.31	1.03
Yang (1979)	1.02	1.12	1.03
Shen and Hung (1971)	0.91	1.18	0.95
Engelund and Hansen (1973)	0.88	1.51	0.96
Ackers and White (1973)	1.28	1.50	1.31
Maddock (1976)	0.99	0.49	0.92

A different picture is painted by the comparative study carried out by Van Rijn (1984b), also using field and laboratory data. The discrepancy ratio, r, defined as the ratio of predicted to observed transport rates in percent were as follows:

	$0.75 \leq r \leq 1.5$				**$0.5 \leq r \leq 2$**				**$0.33 \leq r \leq 3$**			
Data	**1**	**2**	**3**	**4**	**1**	**2**	**3**	**4**	**1**	**2**	**3**	**4**
US Rivers Corps Engrs	53	39	32	6	79	67	61	24	94	87	78	44
Middle Loop River	39	13	37	63	78	37	74	94	96	80	98	100
Indian Canals	30	15	27	3	60	45	48	6	90	73	70	24
Pakistan Canals	23	37	34	13	56	71	71	29	91	94	91	48
Niobrara River Canals	55	13	29	86	95	67	58	98	98	95	98	98
Average of field data	45	32	32	22	76	64	63	39	94	88	84	55
Flumes												
Guy et al., (1966)	40	67	56	68	70	89	85	90	91	98	99	98
Oxford	37	20	31	45	84	38	59	89	96	70	81	96
Stein (1973)	54	73	81	56	70	95	97	97	97	97	100	100
Southampton A	64	49	46	49	85	73	79	82	97	91	94	94
Southampton B	18	12	82	91	81	82	96	97	94	97	100	100
Barton-Lin (1955)	35	60	30	40	65	100	50	65	100	100	100	100
Average of laboratory data	41	46	52	59	77	74	77	89	95	89	94	98
Average of all data	43	37	40	36	76	68	68	58	94	88	88	71

In the above table, column 1 lists values obtained by the method of Van Rijn (1984 a & b); 2 by the Engelund-Hansen formula (1967); 3 by the Ackers-White (1973) formula and 4 by the Yang (1973) formula. The result is poor accuracy by the Yang formula for canals in India and Pakistan, which have the deepest flows of the above data. Since the other formulae produce reasonable results Van Rijn concludes that "the method of Yang must have serious systematic errors at large flow depth. On the average the predicted values are much too small".

4.2.3 Sediment transport of cohesive sediments

Fine sediments (clay and silt fractions) adhere to the same physical processes as coarser particles under saturated, uniform flow conditions and a sediment transport equation can be calibrated to describe their transport. The water viscosity however plays an important role in affecting the settling velocity of the particles.

Non-equilibrium transport of fine sediments occurs, which means that transition to saturated sediment transport capacity conditions is not instantaneous as for coarser fractions, but a time and distance lag is involved. A multi-fraction state of the art model can operate with a traditional equilibrium transport equation for some fractions and with a non-equilibrium formula for others.

The transport of the cohesive sediment in a 2DH model is described by the following unsteady advection-dispersion equation:

$$\frac{\partial hc}{\partial t}+\frac{\partial p'c}{\partial x}+\frac{\partial q'c}{\partial y}=\frac{\partial}{\partial x}\left(hD_{xx}\frac{\partial c}{\partial x}\right)+\frac{\partial}{\partial y}\left(hD_{yy}\frac{\partial c}{\partial y}\right)+E-D \qquad (4\text{-}5)$$

Where:

p',q'	Modified flux field (m^2/s)
c	Concentration (g/m^3)
D_{xx}	Dispersion in the x-direction (m^2/s)
D_{yy}	Dispersion in the y-direction (m^2/s)
E	Erosion function
D	Deposition function

4.2.4 Sediment re-entrainment and deposition

The model should combine theory for non-cohesive and cohesive sediment to model three cases of erosion: cohesive sediment, non-cohesive sediment and a mixture of the two. In the latter case, a linear combination of the cohesive and non-cohesive relationship is used. The erosion rate is calculated assuming both cohesive and non-cohesive sediments, and the actual rate determined via linear interpolation.

A standard cohesive model gives the erosion and deposition functions (E and D):

$$E = E_0\left(\frac{\tau}{\tau_{ce}}-1\right)^m, \quad \tau > \tau_{ce} \qquad \text{(Surface Erosion)} \qquad (4\text{-}6)$$

$$D = w_s c\left(1-\frac{\tau}{\tau_{cd}}\right), \quad \tau < \tau_{cd} \qquad (4\text{-}7)$$

$$E = \frac{\left(C^* - C\right)}{\Delta t}\ ;\ \tau > \tau_{cme} \qquad \text{(Mass erosion, no lag)} \qquad (4\text{-}8)$$

Where:

w_s	Settling velocity, w_s ~ 1 mm/s
τ_{ce}	Critical shear stress for erosion, τ_{ce} ~0.2 N/m^2 for fluid mud and ~0.6 N/m^2 for mud
τ_{cd}	Critical shear stress for deposition, τ_{cd} ~ 0.05 N/m^2
τ_{cme}	Critical shear stress for mass erosion
E_0	Erosion constant, E_0 ~0.1 $g/m^2/s$
m	Exponent (non-linearity) of the erosion, m ~ 1-3
τ	Bed shear stress
C*	Sediment transport capacity concentration

With surface erosion individual particles are removed from the surface at a given rate per square meter, while at higher shear stresses the bed fails deeper as a mass.

4.2.5 Cross-section deformation

Solution of the bed continuity equation determines whether erosion or deposition will occur. Various bed change methodologies can be specified in a 1D model, for example when deposition occurs it is assumed that:

$$dz = aD^b \tag{4-9}$$

with dz the change in bed level, a and b input calibration parameters and D the flow depth.

In 1D models bed changes are usually made uniformly across the section, or in the main channel only.

Minimisation of stream power is another method (1D model) proposed by Chang (1988) in which the combination of channel width and depth gives the smallest energy slope that provides the equilibrium channel shape. However, this method is very sensitive to the sediment transport equation used to determine the energy slope in the minimisation calculation.

Specification of the cross-section deformation is probably the largest drawback of the 1D approach that uses cross-sections, and the 2D models have no such requirement. Yet, in 2D models the number of cells in the main channel should be sufficient to accurately describe the bed and banks.

4.2.6 Hydraulic roughness and its interrelationship with sediment transport

In many models this parameter is fixed on the base of hydrodynamic model calibration, using the Chezy (C) or Manning (n) value, in some cases allowing spatial variation (C or n) in the main channel, banks and floodplain. The hydraulic roughness is however much more important in the modelling of sediment transport than is usually considered and it affects the critical conditions for re-entrainment of sediment and the sediment transport capacity. Temporal variation of bed roughness is sometimes included by empirical functions, but should be refined in future research. However, the fact is that in most analysis re-entrainment conditions are based on a horizontal bed and uniform flow conditions, using the Shields diagram or modified Liu diagram, which is far from reality.

Vegetation often also acts as large-scale roughness. On the river banks it is sometimes partially inundated and during floods it could be completely submerged. Vegetation also undergoes changes with seasons and drought/flood conditions, which makes it difficult to predict the long-term temporal variation of hydraulic roughness.

4.2.7 Coupled solution of flow and sediment equations, with sediment continuity

A coupled solution is required due to the rapidly changing hydrodynamic and sediment transport conditions usually experienced in rivers in Africa. However, coupling in every time step increases computer runtime, and state of the art models can do several morphodynamic time steps before adjusting the hydraulics, especially during slowly changing low sediment transport periods.

4.2.8 Consolidation of cohesive sediment

Modelling of consolidation is important when critical conditions for mass erosion are linked to sediment characteristics and density, and should be considered in lakes, reservoirs and estuaries where sediment densities may vary from 400 kg/m^3 to 1800 kg/m^2.

4.3 Calibration of sediment transport against field data

Considering the requirements of a comprehensive hydrodynamic and sediment transport model, and the many unknowns in the field, it is important that the simulated sediment loads are calibrated against field data. If long term (> 5 years in semi-arid conditions) data are available of total loads at sites along the river, sediment load-discharge or cumulative relationships can be used to compare simulated versus observed data in the calibration process. Verification simulations using single flood and sediment routing through the river should also be carried out. An example of a 1D model sediment concentration calibration and validation on the Berg River, South Africa is shown in Figure 4-3. The prediction seems good but more data are required during floods.

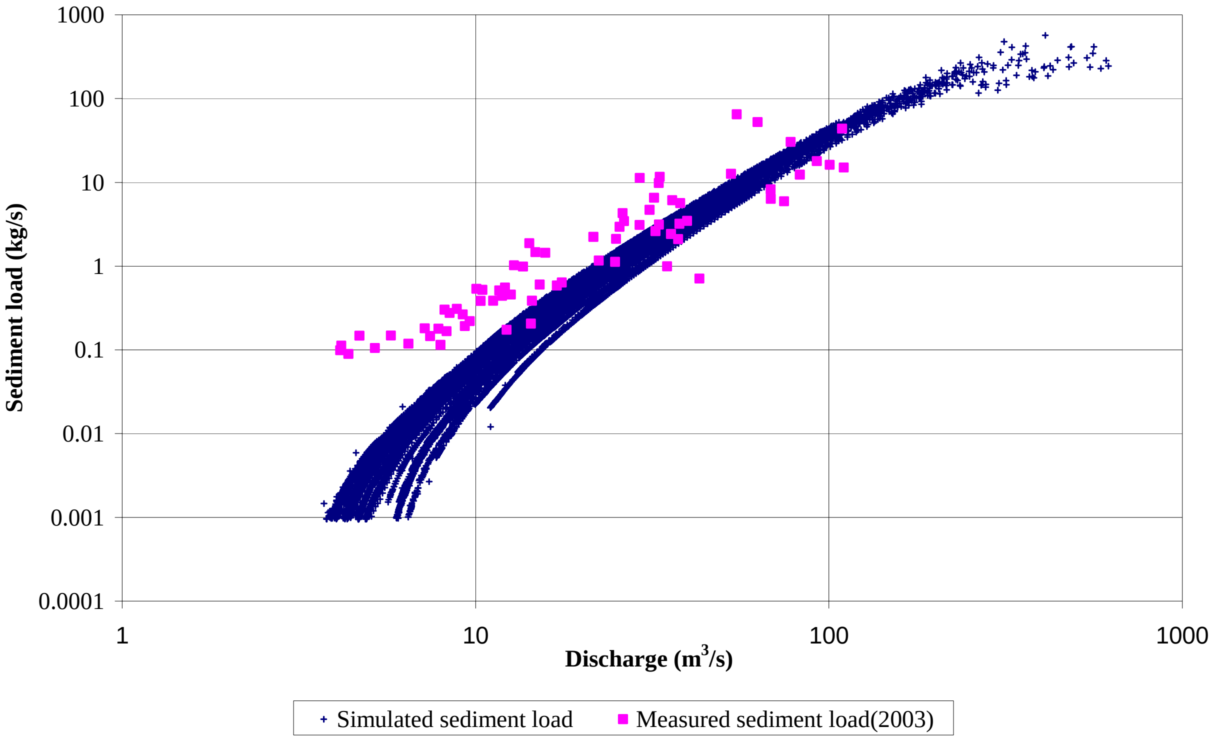

Fig 4-3: Calibrated sediment load-discharge relationship on the Berg River, South Africa

4.4 Model dimensions: Grid spacing and time steps

In order to see 1D and 2D/3D models in perspective, it is important to compare typical grid spacing used. The spacing of course varies with the size, length and of the river to be modelled. For a 1D model of say a 50 km river, cross-sections every 200 m to 400 m are typically considered with about 1 minute to 3 minute time steps. In a 2DH model, the grid spacing across the flow could be anything from 5 m to say 30 m, with longitudinal grid distances the same or a ratio of 2 to 3 larger. In a 2DH approach, the model time steps are usually less than 1 minute in a fully hydrodynamic mode. Two dimensional (and 3D) models therefore run much longer than 1D models with the many grid points to solve in a fully hydrodynamic mode and it is difficult (if not practically impossible) to do long term runs with them where 10 to 20 years have to be simulated.

4.5 Reservoir Sedimentation

In rivers turbulent sediment transport is usually the mode of transport, but in a reservoir or lake the modes of transport could also include colloidal suspension usually considered to be less than 3% of total load, or density current formation due to density differences created by a turbid river flowing into a reservoir.

In relatively shallow reservoirs – say less than 30 m in depth – the mathematical modelling approach described above (2DH) with non-cohesive and cohesive sediments, and bed mass conservation is also suitable for reservoir sediment deposition and re-entrainment simulations for conditions of storage operation as well as flushing of sediments during floods with water level drawdown. Consolidation of sediments and its effect on critical re-entrainment conditions is however more important than in a river.

Density currents (turbidity currents) can be described mathematically by 2DV or 3D models, and have been found to be reliable for field conditions. Some aspects however need further research, such as the erosion and deposition of non-uniform and cohesive sediment, sediment-water layer interaction, numerical modelling of density current formation conditions and density current venting (release through bottom outlets) management.

4.6 Estuary sediment dynamics

2DH models (or 3D) are suitable for simulation of estuary sediment dynamics. Estuaries vary in shapes and sizes and differ from rivers in the following ways:

- The ocean forms an open boundary with a range of tides from neap to spring which affects the flood and ebb tides.
- In regions with a small tidal range (< 2 m) and low river flow, the estuary mouths could close annually during storms at sea. These estuaries breaches naturally when river water spills over the sand bank at the mouth or could be breached artificially in order not to flood low lying developments.
- Ebb and flood tides could follow different flow channels near the mouth of the estuary.
- The ebb tide is often relatively large compared to river inflows.
- Estuaries often have cohesive sediment in the bed.

4.7 Hydraulic Structures

In sediment routing through a river it is important to consider the impacts of hydraulic structures such as groynes, bridges, culverts, weirs, canalised reaches, etc., on the accuracy of sediment transport simulation. Where necessary, 2DH models as described above can simulate complex flow patterns with scour. River meander migration and local bridge constriction scour of 7.5 m depth was simulated for a 2.4 km proposed bridge over the Lower Zambezi River. Figure 4-4 shows a satellite image (2003) which indicates that the upstream meander migrates downstream at a rate of about 85 m/year, while figure 4-5 shows simulated bed levels after a 1:100 year flood. The major flood flow (25000 m^3/s) width in the river is about 7 km wide.

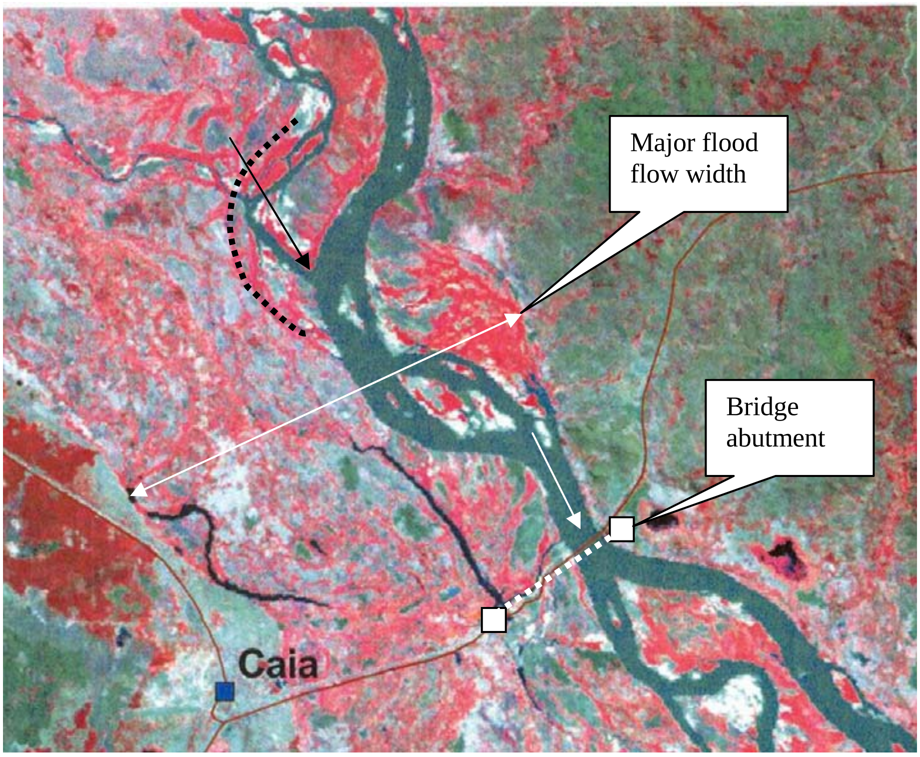

Fig 4-4: Satellite image of Zambezi River at Caia (Dotted white line indicates 2.4 km bridge; Dotted black line shows historical meander)

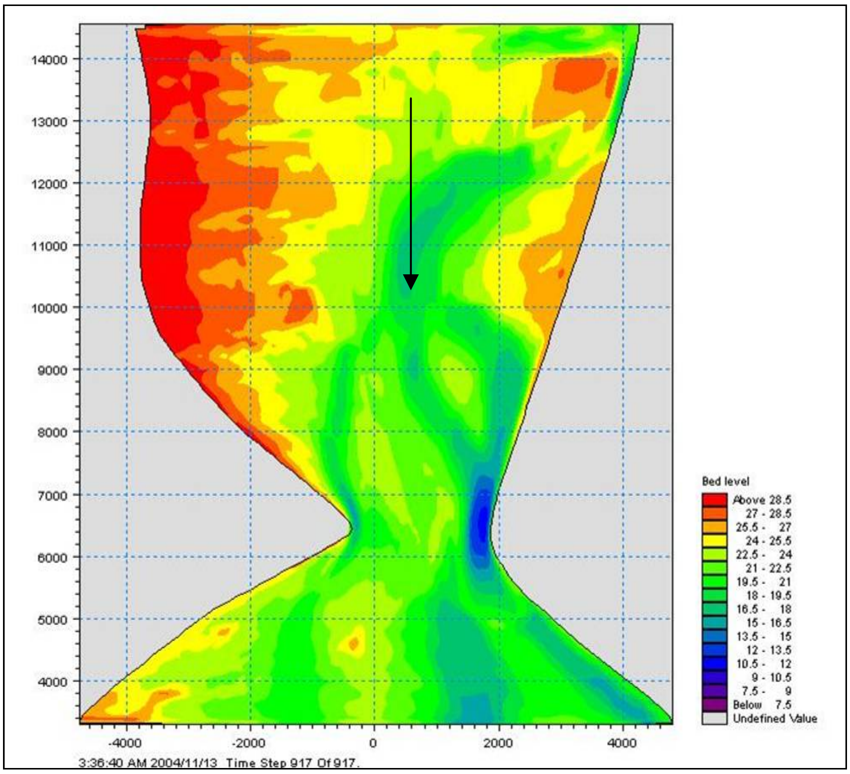

Fig 4-5: Simulated bed level (masl) at end of flood (flood peak 25000 m^3/s, hydrograph duration 2.5 months).

5. SEDIMENT QUALITY

In addition to the prediction of the quantity of erosion and sediment transport, a very important aspect is to assess the role of contaminants attached to the sediment particles, originating from agricultural land or industry. Many sediment yield and river routing models can by now describe the fate of fine sediment that transports the contaminants. Water quality modules and eutrophication modules usually use sediment load data to describe the bio-chemical interactions in the water system related to the contaminants transported by the sediment. However, the sediment transport description is usually very simplified and could be improved in most models. The accurate description of nutrient dynamics in sediments (nitrogen and phosphorus) is very important in the eutrophication models.

6. REMOTE SENSING

Satellite images have been used to quantify soil erosion rates from arid land surfaces in Africa as part of the ESA TIGER Initiative using ENVISAT images. Other useful applications are meander migration rates (Figure 4-4).

Many of the hydrodynamic modelling studies become too expensive because of the cost of surveys which is sometimes 5 times the cost of the modelling. Innovative ways should be sought to reduce this cost. The vertical accuracy of satellite images has improved to about 2 m by using special software to calibrate the images against 1:50000 topographical maps. With Space Shuttle radar images the accuracy is even better at about 1 m. This makes it possible to create DTMs for large areas such as deltas (mostly dry when not in flood) which could be 400 km wide by 200 km long (Zambezi Delta). A laser with GPS and a fixed wing aircraft or helicopter reduces the costs considerably as compared to traditional aerial photography, but is still relatively expensive for large catchments. Only a limited number of laser systems that can scan through the water are currently available. They are affected by high turbidity and their ecological security needs to be verified in rivers.

InSAR (Interferometric Synthetic Aperture Radar) is a new technique which is satellite based and gives 1 to 2 cm height resolution at about 10 m pixel size. Deformation (erosion and deposition) images of up to 100 km x 100 km in size are possible.

In future more research is needed on linking sediment-related variables such as turbidity in a lake/reservoir to other variables such as chlorophyll which can also be remotely monitored, to possibly improve management.

7. CONCLUSIONS AND RECOMMENDATIONS

In large catchments > 2000 km^2, observed data, regional empirical approaches and hydrological routing models incorporating MUSLE to determine the sediment yields are still the most reliable. In smaller catchments investigation of soil erosion, transport and sediment yield could be carried out making use of physically-based mathematical models, of which the “SHE” based model with a 2D approach seems to be the most advanced, but data availability will determine whether other simpler approaches should not be followed.

Sediment transport through a river could be simulated by fully hydrodynamic 2DH model (quasi-3D with the addition of secondary flow at river bends), but it is important to include the description of cohesive sediment re-entrainment and transport, as non-equilibrium and non-uniform transport, especially in lakes and reservoirs. In consequence of the inaccuracy of equilibrium sediment transport equations, mathematical models should be calibrated and verified against field data.

In semi-arid regions the sediment transport capacity of the fine sediment (clay and silt) which is dominant in the water column, far exceeds the actual sediment transport which is limited by the availability from the catchment, and more research is needed in this regard.

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9. GLOSSARY

ACRU	Agrohydrological modelling system of Agricultural Research Unit, South Africa
AD	Advection Dispersion
ANSWERS	Areal Nonpoint Source Watershed Environmental Response Simulation Model
CREAMS	Chemical Runoff and Erosion from Agricultural Management Systems
DTM	Digital terrain model
EUROSEM	European Soil Erosion Model. Department of Environmental Sciences Lancaster University
FSL	Full supply level
GIS	Geographical Information System
LISEM	A Single-event physically-based hydrological and soil erosion model for drainage basins
MUSLE	Modified Universal Soil Loss Equation
NTU	Nephelometric Turbidity Unit
OBS	Optical backscatter Turbidity meter
RUSLE	Revised Universal Soil Loss Equation
SHE	Système Hydrologique Européen hydrological modelling system
SHETRAN	Based on SHE, SHETRAN is a 3D physical based model with coupled water and sediment transport for multi-fractions
USLE	Universal Soil Loss Equation
WEPP	Water Erosion Prediction Project, National Soil Erosion Research Laboratory, USA
2DH	Two-dimensional model in plan, vertically integrated
2DV	Two-dimensional model in the vertical